National Association of Geology Teachers

EASTERN SECTION

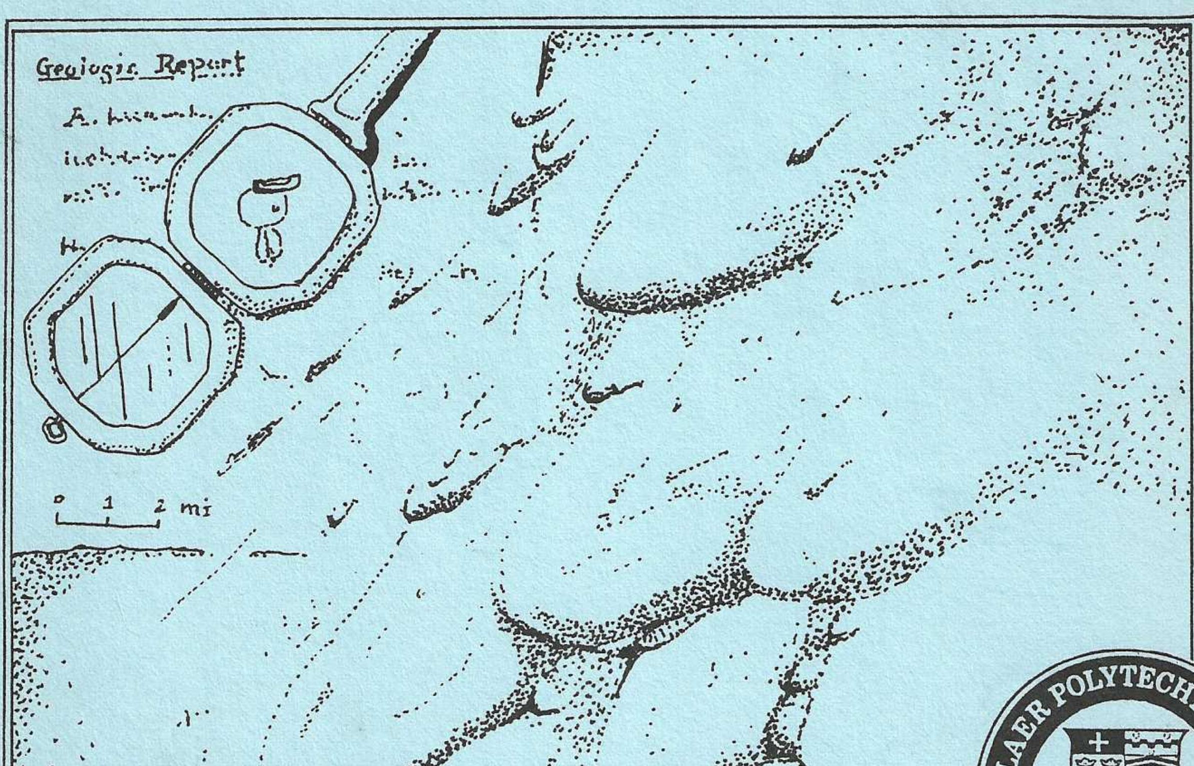
GUIDEBOOK

for FIELD TRIPS

Spring Meeting April 29 - May 1, 1983

Rensselaer Polytechnic Institute

Troy, New York



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PREFACE, WELCOME, AND ACKNOWLEDGEMENTS

Welcome to the 33rd annual meeting of the Eastern Section of the National Association of Geology Teachers. We have arranged for you a fine program of workshops and field trips. Two field trips have been scheduled to classical sites in eastern New York State, one trip will examine bedrock and the second glacial drift.

I wish to extend my thanks to Robert G. LaFleur, Professor of Geology at Rensselaer and leader of one of the field trips and to Paul Stevens, Rensselaer graduate student and coordinator for this meeting. Gloria Hoffman in her usual efficient manner typed this manuscript, and Linda Raine designed the cover.

Enjoy your visit to Rensselaer. Please remember that in many ways the ground on which you tread is hallowed: pioneers of geology, such as Amos Eaton, Ebenezer Emmons, James Hall, Sir Charles Lyell, Douglas Houghton, and numerous other eminent geologists have trod here before you.

Gerald M. Friedman
Editor

REMINISCENCES AND REFLECTIONS OF THE INCOMING PRESIDENT OF THE EASTERN SECTION OF THE NATIONAL ASSOCIATION OF GEOLOGY TEACHERS

by

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This is the 33rd annual meeting of the Eastern Section of NAGT. This meeting here at Rensselaer brings to mind nostalgic memories of my participation at the first meeting of this section on April 20 and 21, 1951 at Lafayette College in Easton, Pennsylvania. I vividly recall presenting a paper at that organizational meeting and was thrilled when the title of my paper was listed in Science and a reprint request arrived from California. As a charter member I enjoyed the experience of being involved in the organization of this section and recall with fond memories my return trip in the car of our sectional and later National president, Chilton (Chip) Prouty. Our second sectional meeting was held at Vassar College. The minutes of the meeting in volume 1 of the Journal of Geological Education (1952, p. 2) list my name as a speaker, but I do not recall what I spoke on. At that meeting doughty Shephard W. Lowman of Rensselaer was elected Vice President, and the third meeting was consequently held here in the halls of Rensselaer. At this third meeting Roland F. Beers, Professor of Geophysics at Rensselaer spoke on "Amos Eaton," pathfinder of North American stratigraphy and founder of Rensselaer. For those who do not remember Roland Beers, he was a Rensselaer graduate of the class of 1921 who distinguished himself in geophysics and whose many achievements included service on committees of the National Academy of Science, as a college president, and as president and director of companies, like the Geotechnical Corporation and his own well-known firm Roland F. Beers, Inc. Other speakers included James R. Dunn, Professor of Economic Geology at Rensselaer, and more recently President of the American Institute of Professional Geologists and recipient of the prestigious Van Couvering Award of the Institute.

Although the meeting which we hosted here at Rensselaer was the third

rather than the first meeting of the eastern section we can be proud to have sponsored the New England section whose first annual meeting was held here in conjunction with our third annual meeting on April 3 and 4, 1953. The first business meeting of the New England section was held here while the members of the eastern section addressed their own business meeting. Shep Lowman became the third president of the eastern section. Lowman by this time had already much of his distinguished career behind him. In my textbook "Principles of Sedimentology," co-authored with John E. Sanders (1978) I report on one of Lowman's contributions (p. 19-20).

> "Beginning in the late 1940's and early 1950's, the first large-scale sedimentological research projects materialized. There had been large-scale research projects before, such as the boring of the Atoll of Funafuti in the Pacific Ocean at the close of the 19th century, but such early efforts were isolated. The 1947 report of the Research Committee of the American Association of Petroleum Geologists, under the leadership of Shephard W. Lowman of Shell Oil Co. and Rensselaer Polytechnic Institute, stated that research in sedimentology is the most-urgent need in petroleum geology. Project 51 of American Petroleum Institute led to a methodical and detailed study of modern depositional environments on a scale not previously attempted. With the aid of research vessels and research teams, modern marine- and deltaic depositional environments were explored. Much of the background of this largest-of-all projects of the American Petroleum Institute was prepared by Lowman who first conceived the idea."

No wonder Rensselaer developed a reputation in sedimentology and petroleum geology.

Of the sectional meetings of the 1950's my next participation was on April 19 and 20, 1957 at Princeton University. By this time I had moved to Tulsa, Oklahoma, and I recall a presentation to the Board of Directors of Stanolind Oil & Gas, now Amoco Production Co., on a problem of carbonate sedimentology and then catching a flight to attend our eastern section meeting. At this meeting Robert G. LaFleur of Rensselaer, who is field-trip leader at this meeting spoke on "teaching elementary historical geology at colleges and universities."

At an informal meeting of the section on December 14, 1957 Joseph L. Rosenholtz of Rensselaer "emphasized that oil companies are very much interested in getting Rensselaer geology Majors even though they have had no graduate work" (Journal of Geological Education, vol. 6, p. 34). In 1960

Rosenholtz became the second Rensselaer president of our section. Since Rosenholtz died in 1963 let me quote from my article "Geology at Rensselaer Polytechnic Institute: an American Epitome" (Northeastern Geology, vol. 3, p. 18-28, 1981).

> "In the 1920's and 1930's the field of sedimentology was mostly concerned with provenance studies. A few species of heavy minerals are diagnostic of a particular kind of parent rock; mere identification suffices to determine provenance. When heavy minerals have been determined from a sample network of regional extent, the distribution of certain species may form a distinct areal pattern. In the subsurface heavy minerals have proved to be a valuable means for distinguishing one sandstone from another in single boreholes and in matching sandstones from one hole to another. Heavy-mineral studies of this type were the dominant line in sedimentology of the 1920's and 1930's. This work closely depended on careful separations of suites of the heavy minerals. At the time heavy minerals were most commonly separated by means of heavy liquids. Yet better methods of separation were needed. Many advances in geology have taken place because some new tool or technique has been invented or improved. With it new analytical results could be obtained. Rosenholtz and (his colleague Dudley T.) Smith realized this. With their publications "Tables and Charts of Specific Gravity and Hardness for Use in the Determination of Minerals" (1931) and especially "The Dielectric Constant of Mineral Powders" (1936) they helped advance early sedimentology. Dielectric separation of heavy minerals, as developed by Rosenholtz and Smith, became an important technique in provenance studies. W.H. Twenhofel, in his influential book "Methods of Study of Sediments" (1941) co-authored with S.A. Tyler, gives much credit to Rosenholtz and Smith (p. 25)."

As incoming president of this section I serve as the third holder of this office from Rensselaer, after Lowman and Rosenholtz, both of whom have died many years ago. Perhaps I may even consider myself the fourth rather than the third NAGT sectional president from Rensselaer. My friend and own Ph.D. student from Rensselaer, Kenneth G. Johnson, now Professor of Geology at Skidmore College, served as president of this section from 1976 to 1977.

Now I would like to digress from sectional to national matters. The forerunner of NAGT, founded in 1938 in the middle west by a small group of geologists, expanded into a national organization on November 10, 1951, at the meeting in Detroit, of the Geological Society of America. There a National Constitution was drawn up and I became an officer of the first

National Council. As Kurt E. Lowe, our first National President, expressed it "the first milestone had been reached," (Journal of Geological Education, vol. 1, p. 1, 1952). My function was that of Treasurer and I recall today with some trepidation that the first expanded cash of NAGT came from my own pocket. I served as National Treasurer until the spring of 1955. The society records indicate my service only through 1954, but a change in calendar year extended my term. Preceding my second year in office (1952-1953) the Nominating Committee selected a double slate and I was put up against Chauncy D. Holmes of the University of Missouri and to my surprise was re-elected. When Freeman Foote of Williams College, Chairman of the Nominating Committee, asked me in 1954 to serve again I had to decline. A growing family forced me temporarily to abandon academe in favor of the flesh-pots of industry.

The office of the National Treasurer in the 1950's included subscriptions and distribution of the Journal of Geological Education, no mean task, which I accomplished with my wife's help. Everything had to be done by hand and was slow and time-consuming. My sense of closeness with the other national officers, especially Ralph Digman and Kurt E. Lowe, made service on the association's council a distinct pleasure. In 1953 when I was on the faculty of the University of Cincinnati Ralph Digman, then national NAGT Secretary, invited me for the winter vacation to Binghamton, N.Y., where he held a faculty position at Harper College. However, on December 20, 1953, prior to my visit, Digman passed away which shocked those of us who held him dear. My appreciation of Kurt Lowe was expressed as his citationist when he received the Neil Miner Award at the NAGT Meeting in Mexico City (Journal of Geological Education, vol. 16, p. 195-196, 1968).

A specific event as a national officer stands out in my mind for the year 1954. Before the annual convention of the Geological Society of America in Los Angeles I received a call from President Leland Horberg requesting me to take his place as NAGT representative at the meeting of the Board of Directors of the American Geological Institute. Horberg pleaded inability to come to Los Angeles, but was not specific why. On my return I discussed with him on the phone details of what had transpired at the meeting. Shortly thereafter I received news of Horberg's death. This was to me a special loss; not only was Horberg my close colleague on the National Council, but I was his student in the field in Wyoming in the 1940's.

As a member of the National Council I became involved in sectional matters. Maps of future sections had been drawn up which presented some difficulty. Until the founding of the Eastern Section the original association was an organization established in the Mid-west. Establishment of the Eastern Section was encouraged by the parent group. When this occurred the original organization found itself automatically demoted to a sectional organization and took the name of Central Section. However, even its area was too large to be effective and the National Organization had drawn up a plan dividing the central section into west central (still termed Central) and East Central. Such a further decimation of the original organization created some resentment. At the 12th annual meeting of the original Association of Geology Teachers, now Central Section, on April 18, 1952 at Kent State Teachers College, Kent, Ohio, in conjunction with the annual meeting of the Ohio Academy of Science I made my plea, as a speaker, for the establishment of the East Central Section. It was a traumatic experience and the next day my dean at the University of Cincinnati called me on the carpet. Yet for the following year I organized the first annual meeting of the East Central Section which was held at my home university, the University of Cincinnati on April 10-11, 1953, and a new section had been born, the second section of which I had become a charter member. This first meeting proved to be a real success; we even appeared on television.

While at Stanolind, now Amoco Production Co., I found that no NAGT section existed in Oklahoma, and it was time to do something about it. At a meeting of the Oklahoma Academy of Science in Norman in 1961 Edward C. Stoever, Jr., of the University of Oklahoma, later National President of NAGT, and I prevailed on prospective members to found a new section which held its first annual meeting on December 8, 1962 at the University of Tulsa. Organization of this section made me a charter member of a third NAGT section. Refreshing my memory of this meeting from the NAGT Transactions (Journal of Geological Education, vol. 11, p. 78, 1963) I had a similar role there as I have at this meeting here at Rensselaer, (1) I became incoming Sectional President and (2) I served as field-trip leader "to examine Pennsylvanian sandstones and carbonate rocks of the Tulsa area, with emphasis on the environment of deposition." On October 19, 1963 the Oklahoma section held an informal meeting. The NAGT minutes quote the sectional secretary (Journal of Geological Education, vol. 12, p. 36, 1964) that after touring "the geologic facilities

of 2 outstanding research laboratories" the group "adjoined to the residence of Dr. Gerald M. Friedman for coffee and cake and an informal get-together. Dr. Friedman showed kodachrome slides of Alaska and described his geological reconnaissance of the state by plane, helicopter and on foot."

On December 7, 1963 the Oklahoma section held its second annual meeting at the University of Oklahoma, Norman, in conjunction with that of the Oklahoma Academy of Science. I presided at that meeting which was my last formal assignment as an officer of NAGT prior to this meeting here at Rensselaer.

As I was also interested in the formation of other sections beyond those three of which I am a charter member I had hopes to attend the first annual meetings of other sections, especially that of Texas. Unfortunately such visits could not be arranged, but on December 8 and 9, 1961 I attended the third annual meeting of the Texas section held in conjunction with the Texas Academy of Science at Galveston, Texas. A highlight of this meeting was a field trip to examine the effects of hurricane Carla, a devastating event. Slides taken on this field trip have travelled with me around the world illustrating the effects of episodic storm events along a shoreline.

In my normal schedule I look forward to what is ahead. My incoming sectional presidency allows me to reflect and reminisce on the past, something I otherwise rarely take time for.

FIELD-TRIP A

DEEP-WATER GRAVITY-DISPLACED DEPOSITS MARGINAL TO THE SHELF EDGE OF THE NOW-VANISHED PROTO-ATLANTIC (IAPETUS) OCEAN

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The fascination of the area around Rensselaer Polytechnic Institute is its diversity of sedimentary geology: in one day of field trips we can examine sedimentary facies of Cambrian-Ordovician age which originated in shallow as well as in deep marine waters. Few other areas can match this diversity of sedimentary facies. The geologic coincidence for this diversity of sedimentary environments in the area of the Rensselaer Polytechnic Institute Campus is its unique location: from Early Cambrian through Early Ordovician R.P.I. would have been on a carbonate shelf. Between Early Cambrian and Early Ordovician times the shelf to basin transition was east of Rutland, Vermont. Tectonic movements shoved Cambrian and Ordovician rocks of slope, rise and basin facies across the shelf facies so that today the exposures on and near the Campus of Rensselaer Polytechnic Institute are basin or basin margin (rise facies with shelf facies of Cambrian and Ordovician age occurring to the west (Friedman, 1972, 1979; Friedman et al, 1982 (Fig. 1).

During Cambrian-Ordovician time, most of the North American continent was a shallow epeiric shelf sea, like the present-day Bahama Bank. At the eastern edge of this shallow sea, i.e. at the eastern edge of this continent, a relatively steep slope existed down which carbonate sediment moved by slides, slumps, turbidity currents, mud flows, and sandfalls to oceanic depths to come to rest at the deep-water basin margin (rise), where a shale facies was deposited (Sanders and Friedman, 1967, p. 240-248; Friedman, 1972, p. 3; Keith and Friedman, 1977, 1978; Friedman and Sanders, 1978, p. 389,392). Shale also formed much of the basinal facies in the deep water beyond. Because allochthonous transport has been inferred for large blocks of rocks presently exposed on and near the R.P.I. Campus, the evidence on the ground shows that the Campus is the site of Cambrian and Early Ordovician rocks of basin margin (rise) and deep basin facies (shales deposited in the Middle Ordovician

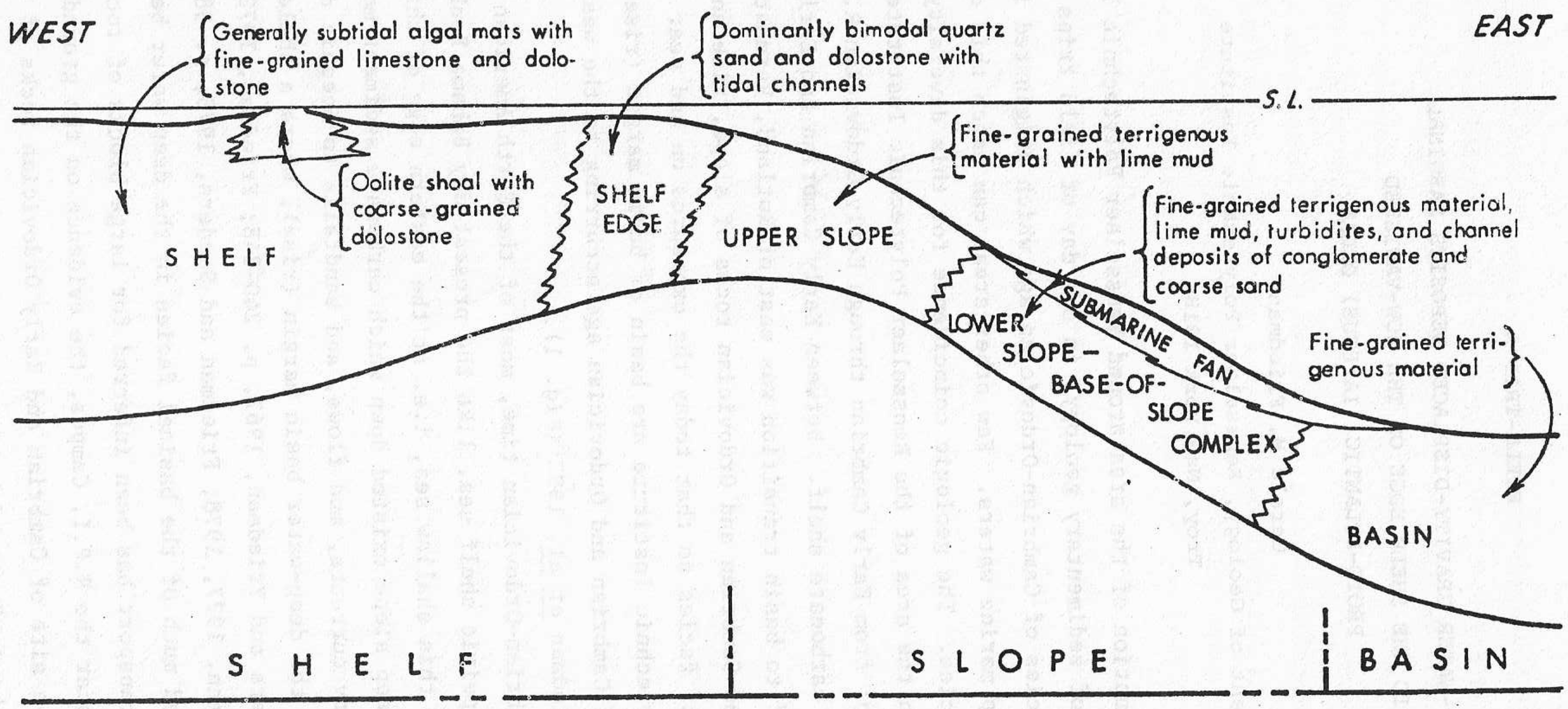

Figure 1. Diagrammatic sketch map showing depositional environments and characteristic sediments of Proto-Atlantic (Iapetus) Ocean for eastern New York and western Vermont during the Early Paleozoic (Keith and Friedman, 1977, Fig. 2, p. 1222; Friedman, 1979, Fig. 1, p. 48).

(Schenectady) west of Campus are autochthonous basin facies). Thus deep-water basin margin (rise) and basinal facies can be visited on and near the R.P.I. Campus, whereas to the west carbonate shelf facies are exposed that are analogous to those of the west shore of Andros Island on the Great Bahama Bank (Fig.1). The paleoslope was probably an active hinge line between the continent to the west and the deep ocean to the east, similar to the Jurassic hinge line of the eastern Mediterranean between carbonate shelf facies and deep-water shales (Friedman, Barzel, and Derin, 1971). Such hinge lines in the early geosynclinal history of mountain belts are fixed by contemporaneous down-to-basin normal faulting (Rodgers, 1968, quoting Truempy, 1960), as probably occurred with the rocks of the area near R.P.I. Later thrusting to lift the deep-water facies across the shelf facies along hinge-line faults resulted in the contiguity of the two facies. This later displacement was so great that the Cambrian and Early Ordovician deep-water sediments were shifted far west of their basin margin.

This field trip has been divided into two parts, each part corresponding to half a day. In the morning and early afternoon we shall study facies of deep-water origin and in the late afternoon those of shallow epeiric origin. Each of the two depositional settings will now be explained.

DEEP-WATER SETTING: A SLOPE-FAN-BASIN-PLAIN MODEL

The strata of deep-water setting are part of the Taconic Sequence (Fig. 3). These rocks have received the attention of geologists for more than 150 years, and because of their exceedingly complex structural and stratigraphic relations have been the object of considerable debate. In fact approximately 150 years ago Ebenezer Emmons' advocacy of the Taconic System (1842, 1844, 1848, 1855) and the division of thought on this problem resulted in the famous duel between James Hall and Emmons which ultimately forced Emmons to leave New York State. A court decision involving several of the most well-known geologists of the last century assured Hall's victory by forcing Emmons out of New York; he settled in North Carolina away from his Taconic rocks.

Strata of the Taconic Sequence extend from north to south approximately 150 miles (Fig. 2) and for the most part within New York State are composed of shales and sandstones. Carbonate rocks are minor by comparison, but are important as they reflect depositional conditions. Although the stratigraphy

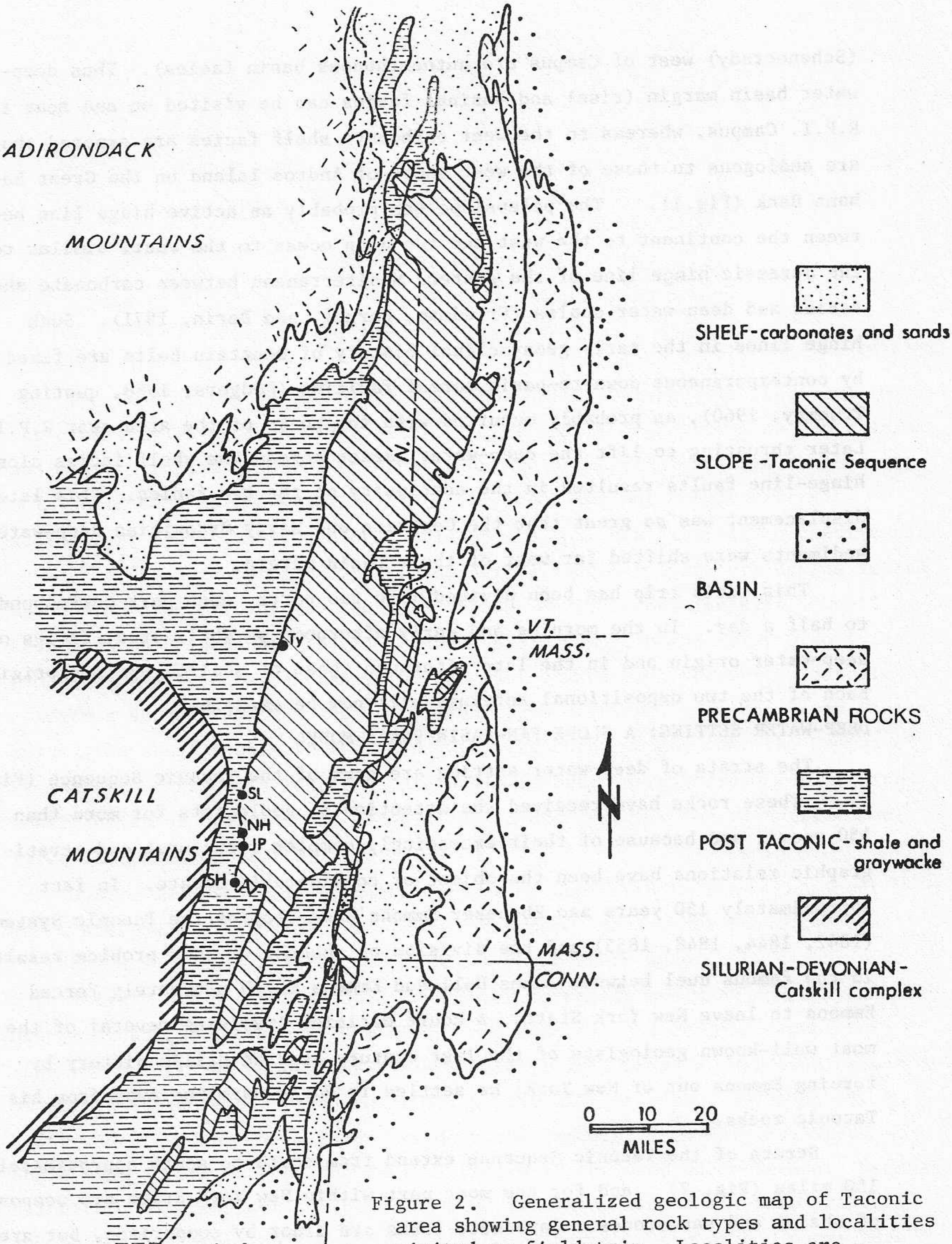

Figure 2. Generalized geologic map of Taconic area showing general rock types and localities visited on field trip. Localities are - SH = South of Hudson (Stop 1); JP = Judson Point (Stop 2); NH = Nutten Hook (Stop 3); SL = Schodack Landing (Stop 4); TY = Troy area (R.P.I. Campus) Stops 5-7) (after Keith and Friedman, 1977, Fig. 1, p. 1221).

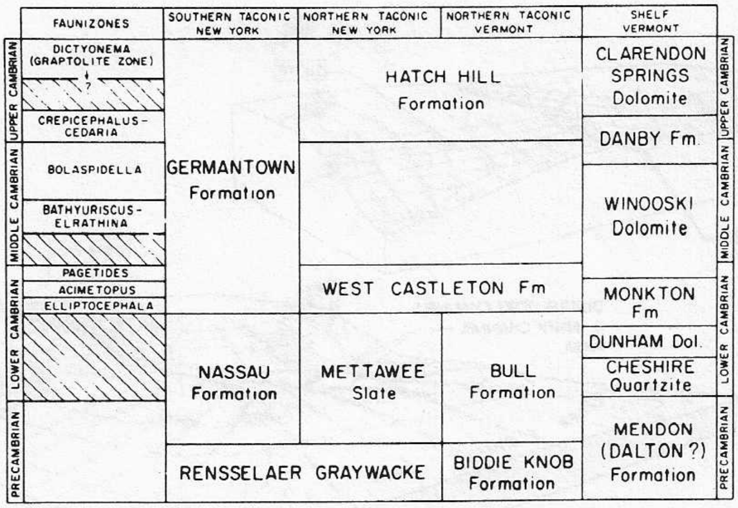

Figure 3. Stratigraphic correlation chart. Deep-water deposits seen on field trip are of West Castleton Formation (Lower Cambrian). Hachured areas represent faunizones not represented in rocks shown on chart (Keith and Friedman, 1977, Fig. 3, p. 1222).

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and tectonics of the area have been the subject of considerable controversy, a debate that has become known as the "Taconic Problem," stratigraphic succession and structure have more recently been clarified (Bird and Rasetti, 1968; Zen, 1967).

Environmental reconstruction for the Cambrian part of the Taconic Sequence in eastern New York State indicates a depositional environment analogous with a modern continental rise or more specifically with a slope-fan-basin-plain model (Fig. 4, Keith and Friedman, 1977, 1978). Carbonate sediment and generally coarse quartz sand were removed from the Cambrian shelf and deposited with muds of the slope, now slates and siltstones, by a variety of processes at work on the slope and within submarine canyons. The shelf-derived sediment can be divided into six main lithofacies, each bearing the imprint of the principal process or processes involved in its deposition. These include: (1) carbonate-clast conglomerates (inferred products of debris flow), (2) massive, coarse sandstones (apparent deposits of fluidized sediment flow and grain flow), (3) graded sandstones and limestones (presumed turbidites), (4) parallel-laminated sandstones and limestones (probable turbidites), (5) thin, structureless micrites (inferred deposits of vertical settling-out of suspension), and (6) current-ripple-laminated limestones and sandstones

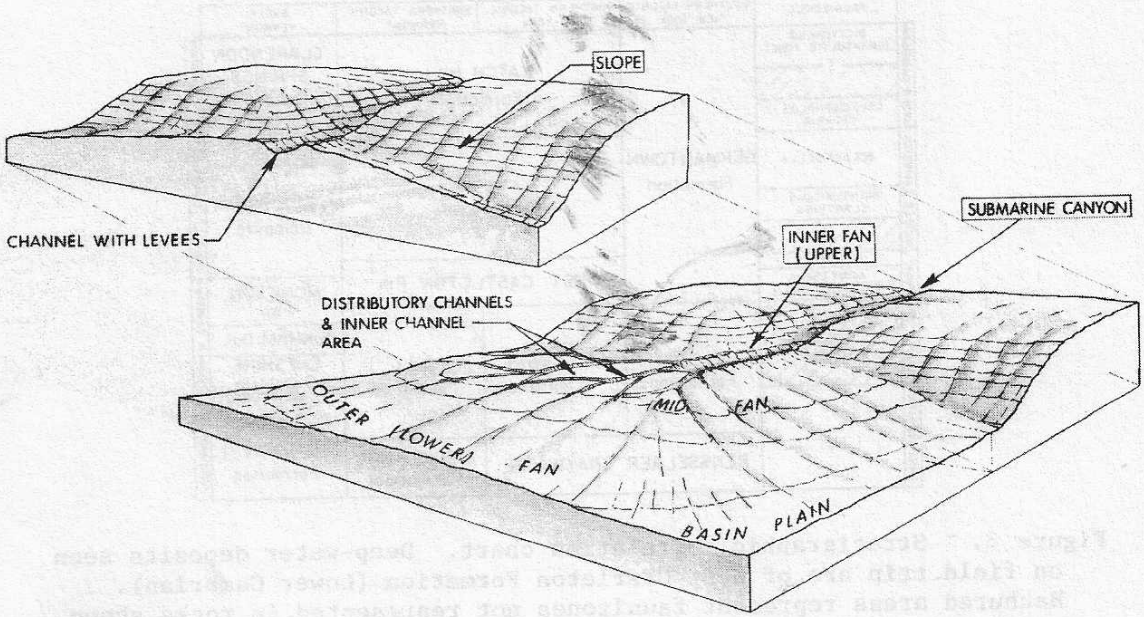

Figure 4. Diagrammatic block diagram of submarine canyon and fan complex, showing major morphologic features. Vertical relief exaggerated (Keith and Friedman, 1977, Fig. 19, p. 233; Friedman, 1979, Fig. 10, p. 63).

(thought to be the products of reworking by contour-following bottom currents or submarine overbank levee deposits). All of these processes were working together or in opposition. Analysis indicates that only the lower slope and base-of-slope portion of the early Paleozoic continental margin has been preserved in the Taconic Sequence (Keith and Friedman, 1977, 1978).

We shall discuss briefly these lithofacies.

Carbonate-Clast Conglomerate (Figs. 5 and 6).

Monomictic carbonate to polymictic carbonate conglomerates occur throughout the Taconic Sequence; a significant percentage of sandstone clasts may be present in some beds. The clasts have a general preferred orientation parallel to the bed boundaries, where they are exposed, but some clasts in a bed will be oriented up to 90° to the general trend.

Conglomerates resembling those described here have been mentioned in the literature extensively (Walker, 1970; Mountjoy et al., 1972; Walker and Mutti, 1973; Walker, 1975; Friedman and Sanders, 1978). Walker (1975, 1976) has proposed descriptive models for conglomerates of turbidite association (re-

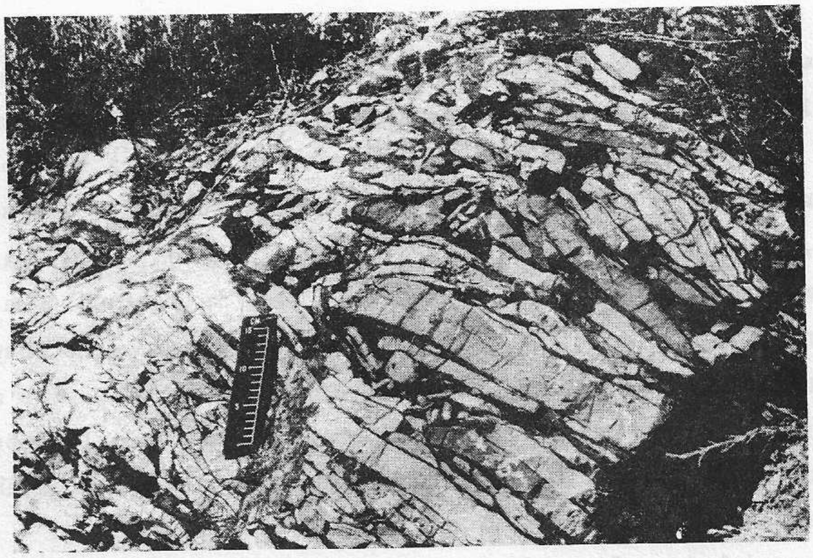

Figure 5. Carbonate-clast conglomerate of shallow-water origin displaced by debris flow from shelf edge into deep water. From exposure south of Hudson (Stop 1 on Fig. 11, SH on Fig. 2).

Figure 6. Rubble of incoherent slump or debris flow composed of boulders of limestone, sandstone, and chert. This rubble, known as brecciola, originated in shallow water behind shelf edge and was displaced into deep-water, dark-colored shales. Note calcite-healed fractures in view. Boulder in center is approx. 30 cm across. Campus of Rensselaer Polytechnic Institute (Stop 5).

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sedimented conglomerates) based on the presence or absence of grading (inverse or normal), stratification and imbrication. The conglomerates seen on this

field trip with their lack of grading and stratification and local imbrication fall closest to Walker's disorganized-bed model. The recognition of a debris-flow model for many of these conglomerates having a lack of organized internal structure has become well established (Dott, 1963; Johnson, 1970; Cook et al., 1972; Hampton, 1972; Middleton and Hampton, 1973; Walker, 1975, 1976; Friedman and Sanders, 1978). A debris flow is defined as a flowing muddy mixture of water and fine particles that supports and transports abundant coarser particles (Friedman and Sanders, 1978, p. 95, 558). The mechanics of motion in any sediment gravity flow are complex, and a simple debris flow model cannot fully explain the features in the conglomerates of the Taconic Sequence (Keith and Friedman, 1977, 1978).

Such a model does fit well with the large clasts in a clay matrix, the lack of size grading, and the poor to nonexistent sorting. The range of composition of the clasts can be easily accounted for, as being derived from the shelf buildup and the basin-margin beds. Some conglomerates appear to be quite local in origin, and interbedded with beds similar to the source beds for the clasts, which also seems compatible with a debris-flow model. The upward decrease in clasts in some beds, with the pervasive preferred orientation and local imbrication all are puzzling as they indicate movement and settling of the individual clasts within the flow. The smaller grain size and presence of some degree of rounding, especially for clasts derived from the shelf, suggest some degree of transport. With increased transport, progressive dilution of the debris flow would take place (as suggested by Hampton, 1972), producing more fluid-like behavior and transition towards turbidity current flow. However, none of the conglomerates discussed here show features indicative of turbidity-current activity such as grading and stratification. At this point all that one can say is that the depositional mechanism appears to be closer to debris flow than to any other (Keith and Friedman, 1977, 1978).

It is not clear whether the conglomerates of the Taconic Sequence were deposited as sheets or were confined to channels. Some of the conglomerates are associated with turbidites, which are generally considered to be confined to submarine canyons or to channels on a submarine fan. Thus, these conglomerates might have been similarly confined.

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In summary, the carbonate-clast conglomerates appear to be the products of deposition by debris flow. The question of whether they are channel deposits or sheet flows is not fully resolved, and both types may well be present (Keith and Friedman, 1977, 1978).

Massive, Coarse Sandstone

These beds of massive sandstone show no bedding, lamination, or grading. The beds seem to fall into two groups, which are: (1) coarse-grained sandstone, and (2) thicker, coarse- to very coarse-grained sandstone. The beds are generally very coarse grained, with no internal features other than a few micrite pebbles. In places the beds contain either micrite pebbles, or wisps that stand out on the weathered surface which appear to be concentrated zones of sand that are more resistant to weathering than the bulk of the bed (Keith and Friedman, 1977, 1978).

The massive beds correspond to beds described extensively from turbidite sequences in the literature (Friedman and Sanders, 1978; Walker, 1967, 1970). Beds generally fitting this description have been called "fluxoturbidites" after the original description by Dzulynski et al. (1959). This term has become of limited usefulness due to the vagueness of the description and resulting misuse. Walker (1970) found, after extensive literature study, that there does exist a facies with certain features including: (1) unusually thick beds; (2) coarse grain size; (3) grading that was repetitive, poor or absent; (4) erosional bases with the finer interbeds being thin, irregular or absent; (5) pebbles commonly present; and (6) tops that may be sharp, rather than gradational. Walker (1970) compared these beds to classical proximal turbidites compiled from the literature and found no significant differences (Keith and Friedman, 1977, 1978).

A depositional mechanism that appears to fit these thick coarse-grained, generally structureless sandstone beds is fluidized sediment flow. This mechanism works when a loosely packed sand is subjected to an initial shock, destroying its fabric, so that water is incorporated and the sand liquifies,

i.e., the grains are supported by excess pore pressure. Since the sand is not sealed, pore fluid loss is rapid, and the flow short-lived. As the pore fluid escapes the viscous properties of the mass disappear and the sediment comes to rest. Because the concentration of sediment relative to fluid is high, features associated with traction deposits, such as different types of lamination, cannot form (Keith and Friedman, 1977, 1978).

Generally, the beds of this lithofacies appear to fit a nebulous category of thick, coarse-grained massive sandstones "proximal" in nature (or possibly channel deposits). They were deposited by one or more processes, involving fluidization of the sediment (Keith and Friedman, 1977, 1978).

Graded Sandstones and Limestones

The graded beds are found associated with beds of other lithofacies. These beds are prominent except south of Hudson, where they are only a minor constituent of the exposed section. Shales are interbedded with this lithofacies at all exposures, except for Judson Point, where sandstone beds are commonly in depositional contact with each other, or with only a very thin shale parting between them (Keith and Friedman, 1977, 1978).

The graded beds range in composition from pure sandstone to limestones, with little or no sand. There are some beds that are half sand and half carbonate. Generally, within one exposure the lithology will be fairly constant. At Judson Point, the beds of this lithofacies are essentially pure sandstone. South of Hudson the beds all contain nearly equal amounts of carbonate and sand. Carbonate is present as rounded intraclasts, individual grains, and as a matrix in the sandy beds. The rounded intraclasts are commonly found near the base of the bed. The intraclasts are composed of pelmicrite, pelsparite or micrite. One intraclast of oomicrite was seen. Sparite and pelmicrite occur as matrix for sandy carbonates (Keith and Friedman, 1977, 1978).

Beds of this lithofacies display many kinds of sedimentary structures. Graded beds, parallel lamination, and cross-lamination (commonly ripple lamination) are all common. Grading takes on several forms in the beds studied. Many beds at Judson Point show delayed grading (Dzulynski and Walton, 1965), where most of the bed is coarse- or medium-grained sand, uniformly distributed, up to the very top, where the bed quickly becomes argillaceous with essentially no intermediate grain sizes. The grading then takes place in a narrow zone

at the top, rather than throughout the bed. Beds at the locality south of Hudson commonly show coarse bimodal sand at the base in a carbonate matrix, with the sand decreasing in amount upward, leaving only the carbonate at the top. This would be a type of discontinuous grading with no medium-grained portion (Keith and Friedman, 1977, 1978).

Parallel lamination is quite common. It appears to be especially well developed in the medium-grained sandstone and the carbonate beds. The laminae are generally less than 1 mm in scale, and in the limestone the lamination is commonly due to fine-grained quartz being concentrated along the laminae. The coarse-grained sandstones, as seen at Judson Point, show only faint lamination, if any at all. Ripple lamination is quite well developed in some beds, but is not common. Not seen elsewhere was larger scale cross-lamination that could be considered cross-bedding in a bed south of Schodack Landing. Many examples of the various internal structures, alone or in combination with others, can be seen (Keith and Friedman, 1977, 1978).

Beds of sand-sized material, displaying grading and lamination in a systematic order (Bouma Sequence) and which are interbedded with basinal shales are turbidites (Fig. 7).

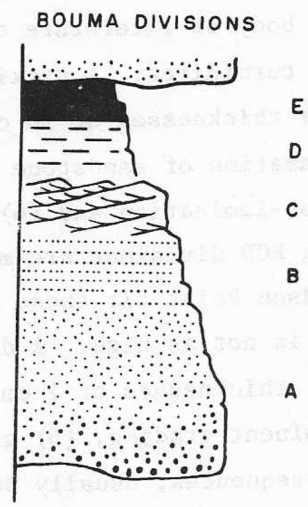

Figure 7. Vertical sequence in sediments deposited by gravity-powered bottom flows. This sequence consists ideally of five divisions, labeled A, B, C, D, E, and named the Bouma sequence after A.H. Bouma (1962):

A. Either a graded sandstone in which the particle size decreases systematically upward or a massive sandstone; the original sand of this "high-speed" depositional layer has a sharp base that divides it from

underlying "low-speed" shaley layer of the preceding sequence. A typically is a product of liquefied cohesionless-particle flow.

B. Parallel-laminated sandstone that represents conditions of upper-flow regime, hence is likewise a "high-speed" structure.

C. Ripple cross-laminated fine or very fine sandstone that represents the lower-flow regime, hence is a "low-speed" structure.

D. Faint parallel laminae of mudstone.

E. Shaley layer at top of sequence. At the contact between E and the overlying sandstone A of the next sequence, abundant sole marks may be present. The fine-grained fallout from the tail of a turbidity current may be difficult or impossible to distinguish from pelagic sediments.

Sequences of turbidites commonly consist of monotonously interbedded alternating and laterally persistent layers of sandstones and shales.

Not all divisions of the Bouma sequence need always be present; sequences may consist of any combination of the five divisions, such as B-C-E, A-E, A-B-C-D-E, E-C, or others. The characteristics of gravity-powered bottom flows include (1) sharp base with sole marks, (2) divisions of Bouma sequence, (3) graded layer or massive sandstone, and (4) monotonously interbedded alternating and laterally persistent sandstones and shales (After Bouma, 1962; Walker, 1976, Fig. 1, p. 26; Friedman and Sanders, 1978, Fig. 12-52, p. 393).

There is a considerable body of literature on the problem of proximal versus distal environments in turbidites. A proximal turbidite has: (1) sharp and flat based beds, (2) thicknesses of 10 cm to 1 m, (3) a sand/shale ratio of about 5:1, (4) amalgamation of sandstone beds present, but uncommon, (5) uncommon parallel and cross-lamination and (6) the typical bed is an AE sequence where the intervening BCD divisions are missing (Walker and Mutti, 1973). The graded beds at Judson Point fit these characteristics fairly well, although the sand/shale ratio is not as high. A distal turbidite has (1) sharp and flat based beds, (2) thicknesses of 1 cm to 10 cm, (3) a sand/shale ratio of 1:1 or less, (4) prominent grading, (5) grain sizes from fine sand to silt, and (6) base-cut-out sequences, usually BCDE, BDE, and CDE (Walker and Mutti, 1973). The graded beds seen at localities other than Judson Point generally fit the first four criteria fairly well, but are coarser grained and generally do not have base-cut-out sequences with the possible exception of certain laminated beds to be discussed in the next section. In general, these graded beds bear more resemblance to distal, rather than proximal, turbidites, but may be transitional (Keith and Friedman, 1977, 1978).

Parallel-Laminated Sandstones and Limestones

Beds identified as belonging to this lithofacies comprise a significant amount of the lithofacies at all of the major sections to be seen on this field trip and are the major lithofacies at Nutten Hook. They are also the only lithofacies besides the conglomerates found in the city of Troy area, especially on and near the R.P.I. Campus.

The beds of this lithofacies range from medium-grained, parallel laminated sandstones (60%), to medium-grained sandstones with parallel lamination and some cross-lamination (22%), to limestones (9%), and coarse-grained sandstones (9%). Most of the coarse sandstones occur at Judson Point. The limestone beds are pelmicrites with the lamination due to the concentration of fine quartz sand and silt along the laminae. In places a bed will contain fossil fragments. Most of the sandstone beds are composed of medium-grained quartz sand with a variable amount of carbonate matrix forming the laminae. Some of the sandstone beds will contain fossil fragments, and, in fact, nearly all the identifiable trilobite fauna recovered by Bird and Rasetti (1968) from Judson Point, and Nutten Hook, and used by them for dating, came from beds identified in the Keith and Friedman (1977) study as belonging to this lithofacies. All but one of the sandstone beds and all of the limestone beds of this lithofacies show lamination of some sort. Commonly, only parallel lamination is present in the sandstones, but some sandstone beds and most of the limestone beds show some cross-lamination.

The beds here probably represent channel-edge equivalents of the coarser, probable channel deposits represented by the conglomerates, massive sandstones and turbidites. For the most part, the beds of this lithofacies would appear to be single beds of division B of the Bouma Sequence.

In summary, beds of this lithofacies are intimately associated with turbidites and may even be types of turbidites themselves (Keith and Friedman, 1977, 1978; Friedman, 1979).

Thin Structureless Micrites

This lithofacies is composed of beds of dense, texturally simple micrite that does not show any features in thin section other than neomorphism where the original lime mud has become recrystallized (Fig. 8). Beds of this lithofacies are found at several of the sections seen on this field trip and comprise a significant amount (approximately 20% of all the lithofacies south

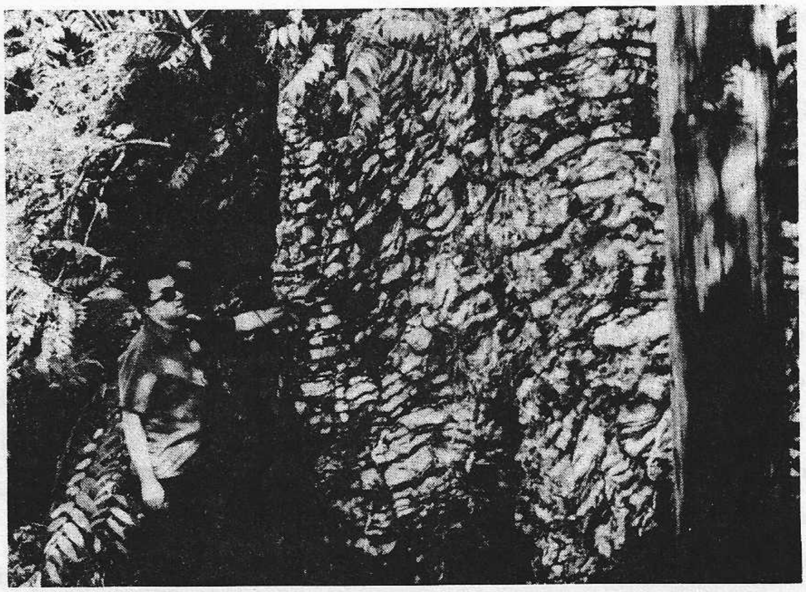

Figure 8. Interbeds of micrite with shale partings. The abundant lime mud was derived from very shallow-water (< 30m) environments. Currents moved the lime mud into deep water. On left is the well-known Polish geologist S. Dzulynski whose work on deep-water deposits is now classical. Schodack Landing (Stop 4 on Fig. 11; SL on Fig. 2.)

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of Schodack Landing and Nutten Hook. Single and multiple beds are found interbedded with beds of other lithofacies. At other localities isolated stringers can be found in places. South of Hudson and Nutten Hook even beds of micrite are interbedded with shale (Fig. 9) and beds of cross-laminated and parallel-laminated pelmicrite. These beds show some pull-apart or boudinage structure and, locally, slump folds. South of Schodack Landing the micrite beds are not associated with any coarser beds and make up 70-75% of that part of the section. Pull-apart is common in these beds. The beds south of Schodack Landing change upward to lenses and stringers of micrite in shale gradually becoming thin and discontinuous. A local conglomerate is present in part of the exposure at Schodack Landing (Keith and Friedman, 1977, 1978).

Thin interbeds of fine-grained limestone (usually micrite) intercalated with dark shale, as described for this lithofacies, have been noted from a number of areas (Sanders and Friedman, 1967; Wilson, 1969). These beds are generally referred to as hemipelagic, because they are a combination of terrigenous sediment and pure pelagic sediment.

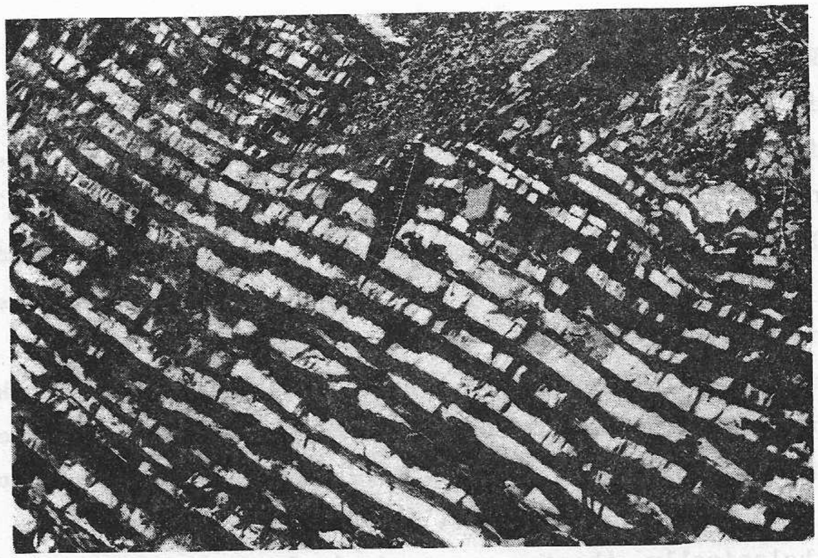


Figure 9. Interbeds of micrite and shale. This micrite is mostly a pelmicrite. The original lime mud was derived from very shallow-water environments. South of Hudson (Stop 1 on Fig. 11 SH on Fig. 2).

The beds discussed here do not contain pelagic microfauna, thus, the only source of abundant lime mud is very shallow water (< 30m) environments. Shallow-water production of lime mud can be quite high, coming from a variety of sources. Once produced, currents can move the lime mud quite easily from the shelf into deeper water. This process would seem to be the only plausible explanation for the micrite beds of this lithofacies. The lime mud was probably carried in dilute suspensions, either by contour currents, nepheloid layers, or dilute turbidity currents (Walker and Mutti, 1973). Isolated beds of micrite could conceivably be attributed to single or an episode of several dilute clouds of lime mud being carried into deeper water. A problem arises, however, with rhythmic succession of micrite and shale beds of very constant and even thickness. Sanders and Friedman (1967) stated that the environmental interpretation of such sequences may be extremely difficult, for similar sequences are seen in near-shore environments.

To summarize, these beds are composed of structureless micrite that is commonly associated with beds or laminae of pelmicrite. This fine-grained material was picked up in suspension on the shelf and carried into deeper water, possibly by contour currents, nepheloid layers, or dilute turbidity currents. The resulting deposits are thin beds of carbonate interbedded with fine-grained terrigenous material (Keith and Friedman, 1977, 1978).

Current-Ripple-Laminated Limestones and Sandstones

This facies consists of thin beds generally ranging from 1.3 cm to 10 cm, with the average thickness of 4.1 cm. This average is considerably less than that for the graded beds (about 19 cm) and less than that for the parallel-laminated beds (about 7 cm), both of which are similar in terms of sedimentary structures. As with the graded sandstones and limestones and thin, structureless micrites, the best exposures of these beds are in the cuts, south of Hudson, Judson Point, and Nutten Hook. No beds of this lithofacies were found south of Schodack Landing. South of Hudson and at Nutten Hook, beds of this lithofacies occur with the thin structureless micrites (Keith and Friedman, '77, '78).

Lithologically, these current-ripple-laminated beds seem to fall into two types -- pelletal limestone with fine-grained quartz sand, or fine-grained quartz sandstones to siltstones. These beds are always laminated with either parallel laminae or striking cross-laminae.

These current-ripple-laminated beds do not show the variety or orderly sequence of features associated with average turbidites. They are also finer grained and thinner bedded than the graded beds described earlier. The laminated beds of this lithofacies probably were deposited by one of three separate processes: submarine overbank levee deposits associated with turbidity currents, distal turbidites, or possibly contour-following bottom currents (Fig. 10). For a more detailed discussion see Keith and Friedman (1977, 1978).

Environmental Reconstruction for Deep-Water Deposits

The rocks of the Taconic Sequence studied here are clearly the products of deposition in a slope environment (Figs. 1 and 4). The depositional mechanisms that were active (debris flow, sediment flow, turbidity currents, hemipelagic sedimentation, and contour currents) appear to be characteristic of the lower part of the slope and the base of the slope. All of these processes, except the contour currents, form a continuum such that one sediment gravity flow could act as a debris flow, sediment flow, turbidity current, or suspended cloud depending upon its time and spatial position on the slope.

There are many problems associated with the reconstruction of the slope environment of the Taconic Sequence. Foremost is the tectonic complexity that has been superimposed since deposition of the sediments. Details of physiography cannot be compared with modern slopes, but the general type and

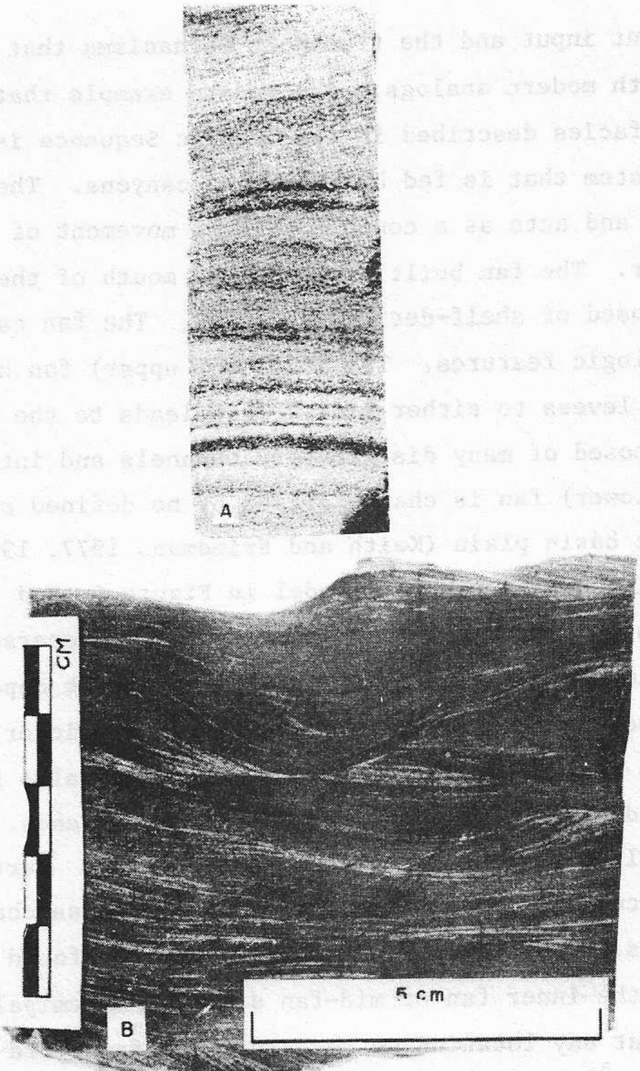


Figure 10. Photographs of modern- and ancient contourites.
A. Core of modern contourite raised from Caicos Outer Ridge, Bahamas, western Atlantic Ocean. This sediment is a well-sorted medium-grained skeletal sand; note horizontal laminae (E.D. Schneider).
B. Polished slab of inferred contourite of Cambrian age (West Castleton Formation) sampled near campus of Rensselaer Polytechnic Institute, Troy, New York. This inferred contourite is a current-ripple cross-laminated pelletal limestone; quartz silt accentuates the laminae. (B.D. Keith.)
(Friedman and Sanders, Fig. 12-51, p. 392)

rate of sediment input and the transport mechanisms that were active can be compared with modern analogs. The modern example that fits very nearly with the lithofacies described in the Taconic Sequence is the slope-fan-basin-plain system that is fed by submarine canyons. The canyon is incised into the slope and acts as a conduit for the movement of shelf sediments into deep water. The fan built out from the mouth of the canyon is then primarily composed of shelf-derived sediment. The fan can be divided into several morphologic features. The inner (or upper) fan has one major channel with prominent levees to either side. This leads to the mid-fan area (or supra-fan) composed of many distributary channels and interchannel areas. The outer (or lower) fan is characterized by no defined channel system and merges into the basin plain (Keith and Friedman, 1977, 1978).

The relationship between the model in Figure 4 and the lithofacies described in this guidebook can be put together. The coarsest and least structured sediments (conglomerates) would form thick deposits at the base of the slope, possibly represented by some of the thicker conglomerates in the area of the R.P.I. Campus. Conglomerates would also be found in the lower canyon and inner fan, associated with coarse sands, as products of debris flow and fluidized sediment flow, respectively. Farther out into the fan, turbidity-current deposition becomes dominant, as channel and interchannel deposits. Overbank levee deposits could be found associated with any channel in the inner fan or mid-fan area. The hemipelagic micrite beds could be found at any location on the slope and fan where they were not subsequently destroyed by current activity. Contourite beds could also be found at any location, depending upon the position of the current at any particular time (Keith and Friedman, 1977, 1978).

The test of this model is whether it can be used to explain some of the exposures seen on this field trip. The exposed sections (South of Hudson, Judson Point, Nutten Hook and Schodack Landing) would best serve to illustrate the application of the model. The first example is the section south of Schodack Landing, shown in Figure 15. The base of the section is composed of considerable thickness of shale, and at least one conglomerate bed; overlying the shale is a sequence of thin micrite and shale beds, possibly the product of transport from the shelf. The conglomerate overlying these beds is probably the result of local slumping, since the clasts all

appear to be derived from the underlying limestone beds. The next higher conglomerate indicates that feeding from the shelf has started, producing coarse sand and biosparite clasts, but only as an isolated event. However, the subsequent presence of channel and interchannel turbidites and a conglomerate, followed by massive sandstones and more turbidites shows active feeding from the shelf and fan development. The beds show definite mid-fan development and possibly an inner fan channel as well. Abruptly, the system appears to have been abandoned, as shown by the resumption of shale deposition, with only a local thin limestone (Keith and Friedman, 1977, 1978).

The section of Judson Point (Fig. 13) is dominated by turbidite beds and several massive coarse sandstone beds. The presence of several AE beds of the Bouma Sequence is characteristic of "proximal" turbidites. The massive beds are common in the lower part, but uncommon in the upper part of the section. The presence of thick massive sands and local conglomerates and thin-bedded laminated sands suggests channel and levee deposits of the inner fan area with the main channel periodically changing its course. Two of the conglomerates are somewhat lenticular in nature and truncate some underlying bed, suggesting channel deposition. For reasons that are not entirely clear, the only carbonate present is a 2 m-thick section at the top of the lower half of the exposure. The beds appear to be fine- to medium-grained hemipelagic and contourite beds, with a conglomerate near the top that contains clasts of the beds below. Possibly the section at Judson Point was influenced by locally dominant sand source. A more likely possibility is that the inner fan area is generally characterized by sands and that carbonates are generally carried farther out by more mature turbidity currents (Keith and Friedman, 1977, 1978).

The section at Nutten Hook is more difficult to interpret, because it is faulted in several places (Fig. 14). The interval below the covered zone is quite sandy and probably represents an environment similar to that just discussed for Judson Point. The section above this zone is dominated by thin interbeds of limestone of hemipelagic type. The lower fault interrupts the section, but the rocks above and below are quite similar. In places thicker laminated beds are distributed that are probably a "distal" turbidite. This was a local area of quiet sedimentation, with virtually no interruption by sediment gravity flows. The section above the upper fault contains several

carbonate conglomerate beds in which the thin limestones, and the top of the section contains conglomerates, thick, coarse, laminated sandstones, and graded sandstones. It would appear that the environment shifted at some point in a "proximal" direction, with the influx of a considerable amount of coarser-grained debris. This shift might be due solely to reactivation of a channel system that was not in use during deposition of the lower part of the section, or due to the buildup of a new channel system, probably in the mid-fan area. The alternative is that the sediments above the fault were brought in tectonically from a more proximal area (Keith and Friedman, 1977, 1978).

The section south of Hudson, New York, contains the highest percentage of shale of any of the exposures (Fig. 12). It is characterized by intermittent, thin turbidite beds, most of which contain coarse sand and even micrite pebbles in the basal portions. Laminated and current-ripple-laminated probable "distal" turbidites also are common and in places associated with thin micrite beds. The middle of the exposure contains a regularly bedded sequence of these two types. The uppermost part of the section is marked by a thin conglomerate bed. This section is more difficult to interpret. It was a site of only intermittent coarse sedimentation, possibly in the mid-fan area. Hemipelagic beds of alternating limestone (micrite) and dark shale are present at the section south of Hudson; currents moved lime mud and terrigenous mud from the shelf into deep water (Keith and Friedman, 1977, 1978).

ITINERARY

Figure 11 is the road log.

Depart from Rensselaer Polytechnic Institute, take People's Avenue west past Samaritan Hospital (on right) downhill to Eighth Street.

Mileage Between points	Cumulative	
0.7	0.7	Proceed for one block to Federal Street, turn right and cross bridge across Hudson River and continue to Interstate 787 south.
0.5	1.2	Enter Interstate 787 south.
4.6	5.8	Take Interstate 90 east; cross Hudson River.
13.2	19.0	Take Exit 12 (sign U.S.9 south) and follow route to Hudson.
1.4	20.4	Enter Columbia County.
3.3	23.7	Junction with Route 9H; continue on U.S.9.
0.6	24.3	Enter Valatie.
1.3	25.6	Enter Kinderhook.
2.1	27.7	Town of Stuyvesant
1.6	29.3	Enter Stuyvesant Falls.
0.7	30.0	Enter Columbiaville; Junction with Route 9J.
4.6	34.6	Enter Stottville.
0.8	35.4	Town of Greenport
0.6	36.0	Enter Hudson; continue south on U.S.9 through Hudson.

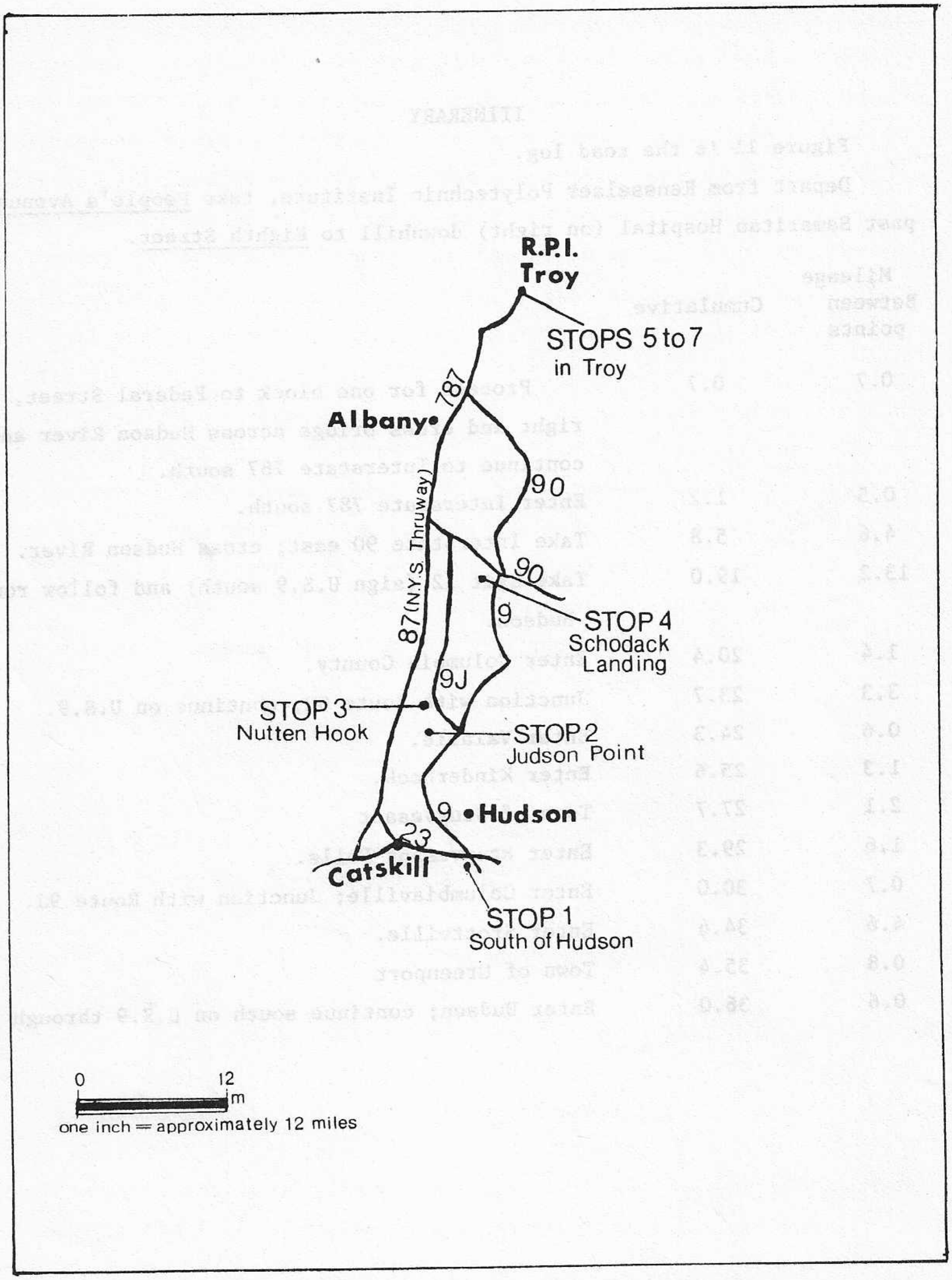


Figure 11. Road log with stops in area of Rensselaer Polytechnic Institute.

Mileage Between points	Cumulative	
5.4	41.4	Junction with U.S. 23; continue south on U.S.9.
0.8	42.2	STOP 1 EXPOSURE ON EAST SIDE OF U.S. 9 (across from a white house) (SECTION SOUTH OF HUDSON)

Figure 12 shows and describes the lithofacies exposed at this stop, and interpretes its depositional setting. The rocks are of deep-water mid-fan origin (Figs. 1 and 4). Note especially the interesting "hemipelagic" interbeds of fine-grained limestone (micrite) and dark shale (Fig. 9) and the carbonate-clast conglomerate (Fig. 5). Read carefully the material covered under the heading of "Deep-Water Setting: a Slope-Fan-Basin-Plain Model" so that you understand the objective of making this stop.

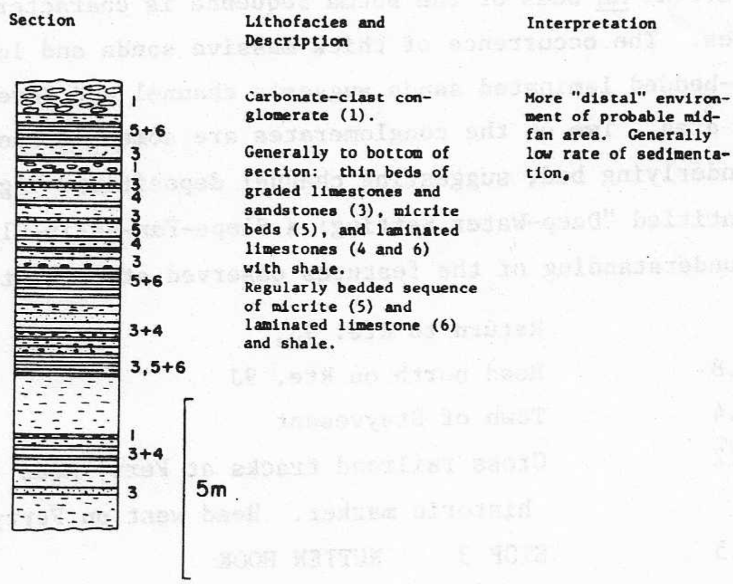


Figure 12. Section south of Hudson Stop 1 (SH on Fig. 2) (Keith and Friedman, 1977, Fig. 23, p. 1237).

		Make a U-Turn and head north on U.S.9
0.8	43.0	Junction with U.S.23; continue north on U.S.9

Mileage Between points	Cumulative	
3.4	46.4	Enter Hudson; continue through Hudson north on U.S.9.
3.7	50.1	Enter Stottville.
2.2	52.3	Enter Columbiaville.
1.2	53.5	Junction of U.S.9, Route 9J, and an unmarked asphalt road identified by sign "Dead End." Turn sharp left onto asphalt road and head west towards Hudson River.
1.0	54.5	At fork take lower road marked "Dead End."
0.1	54.6	STOP 2 JUDSON POINT

Figure 34 illustrates the section seen at this stop and describes the lithofacies; it also provides an interpretation of the depositional setting. The rocks are of inner- to mid-fan origin (Fig. 4). The section is dominated by turbidite beds, and several massive coarse sandstone beds. The presence of several AE beds of the Bouma Sequence is characteristic of proximal turbidites. The occurrence of thick massive sands and local conglomerates and thin-bedded laminated sands suggests channel and levee deposits of the inner fan area. Two of the conglomerates are somewhat lenticular and truncate an underlying bed, suggesting channel deposition. Again review the section entitled "Deep-Water Setting: a Slope-Fan-Basin-Plain Model" for a better understanding of the features observed at this stop.

		Return to Rte. 9J.
1.2	55.8	Head north on Rte. 9J
1.6	57.4	Town of Stuyvesant
0.8	58.2	Cross railroad tracks at Ferry Road (on left); note historic marker. Head west on Ferry Road.
0.3	58.5	STOP 3 NUTTEN HOOK

Figure 14 gives details on this section, including an interpretation of depositional environment.

		Return to Rte. 9J.
0.3	58.9	Head north on Rte. 9J
1.9	60.8	Enter Stuyvesant

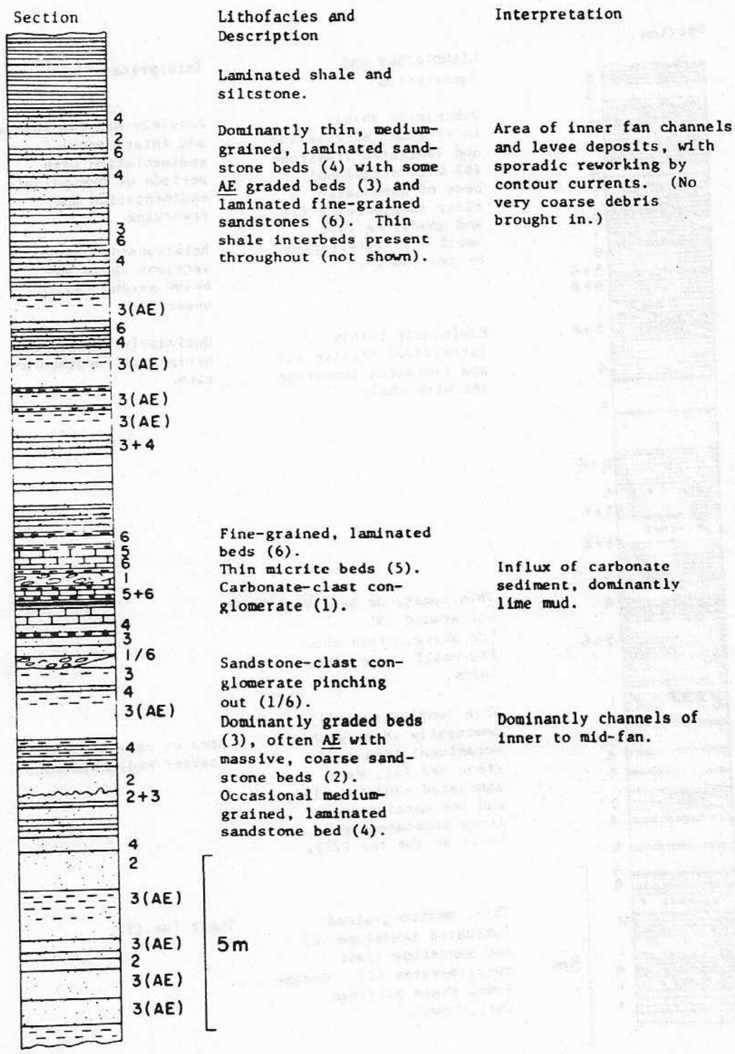
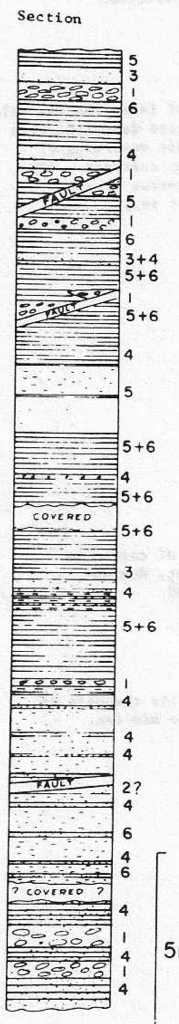


Figure 13. Section at Judson Point (Stop 2; JP on Fig. 3.)
(Keith and Friedman, 1977, Fig. 21, p. 1235).



Lithofacies and Description	Interpretation
Dominantly thinly interbedded micrite (5) and laminated limestone (6) with occasional beds of carbonate-clast conglomerate (1) and sandstone (4). Section is interrupted by two faults.	Possibly mid-fan channel and interchannel sedimentation with periods of hemipelagic sedimentation and reworking. Relationship between sections above and below second fault uncertain.
Dominantly thinly interbedded micrite (5) and laminated limestone (6) with shale.	Dominantly an area of hemipelagic sedimentation.
Thin sandstone beds (4), one graded (3). See above, often showing small scale slump folds.	
Thin conglomerate (1). Dominantly shale with occasional thin sandstone bed (4), rare laminated sandstone (6), and one massive sandstone truncated by a fault at the top (2?).	Area of only sporadic coarser sedimentation.
Thin, medium-grained, laminated sandstone (4) and sandstone-clast conglomerates (1). Occasional shale partings (not shown).	Inner fan (?).

Figure 14. Section at Nutten Hook (Stop 3; NH on Fig. 23) (Keith and Friedman, 1977, Fig. 22, p. 1236).

Mileage Between points	Cumulative	
1.7	62.5	Overpass, New York Central Railroad
1.0	63.5	Railroad overpass
1.9	65.4	STOP 4 EXPOSURES SOUTH OF SCHODACK LANDING

Bus parks on dirt road east of highway. We shall first examine the road cut on the east side of the highway and then walk on dirt road across the railroad tracks to view more fine exposures.

Figure 15 describes and illustrates the section seen and provides interpretation of the depositional setting. Note exposures of bedded micrite with shale partings. This micrite represents lime mud which was derived from the shelf and probably settled from suspension. The conglomerate overlying the bedded micrite is probably the result of slumping, since the clasts appear to be derived from the underlying limestone beds. The strata at this stop show mid-fan development and possibly an inner fan channel as well. Again review the earlier section entitled "Deep-Water Setting: a Slope-Fan-Basin-Plain model" for additional interpretation of depositional setting.

		Continue north on Rte. 9J.
0.9	66.3	Enter Rensselaer County.
0.5	66.8	Enter Schodack Landing.
2.3	69.1	2 overpasses; New York Thruway and railroad
1.5	70.6	Enter Village of Castleton on Hudson.
7.3	77.9	Overpass of U.S.9; junction with U.S.9; follow U.S.9 west (labelled north).
0.2	78.1	Enter Rensselaer.
0.4	78.5	Enter bridge to cross Hudson River.
0.7	79.2	From bridge take Interstate 787 north.
6.8	86.0	Take Troy-U.S.7 Exit (23rd Street, Watervliet, Green Island).
0.5	86.5	Cross Hudson River bridge to Federal Street, Troy.
0.6	87.1	Proceed east uphill on Federal Street to '87 Gymnasium of R.P.I. Campus.

STOP 5 RENSSELAER POLYTECHNIC INSTITUTE, '87 Gym. Exposure behind fence adjacent to gym.

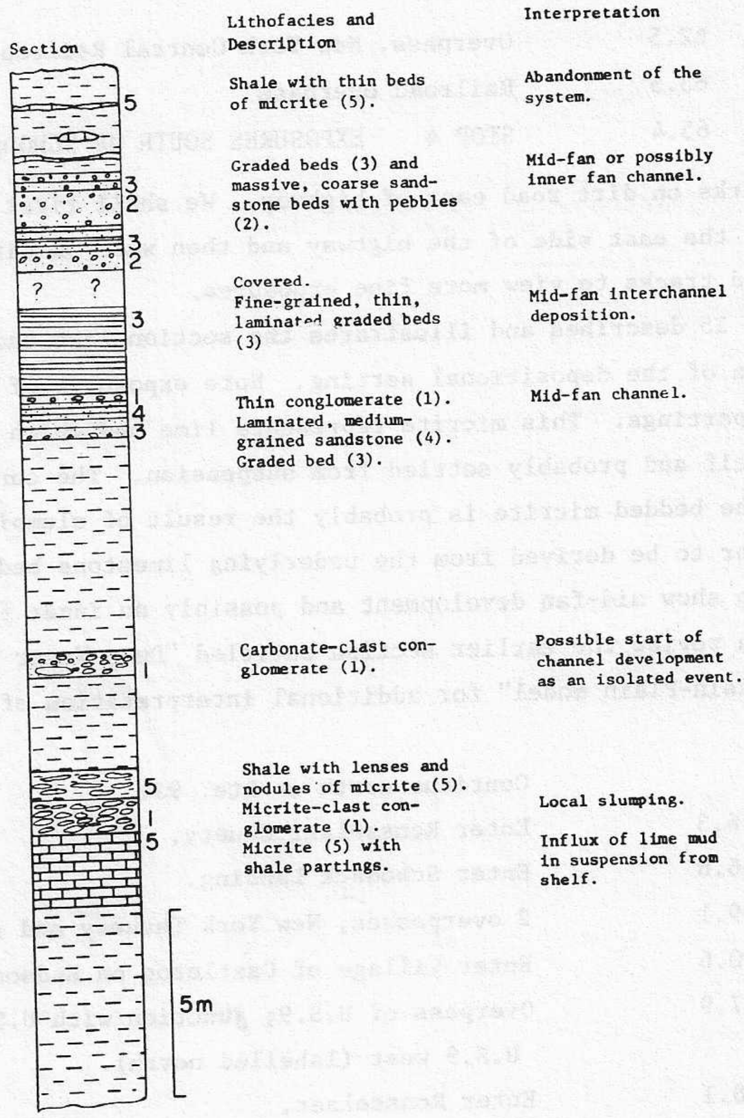


Figure 15. Section at Schodack Landing (Stop 4; SL on Fig. 2)
(Keith and Friedman, 1977, Fig. 20, p. 1234).

Ruedemann (1930, p. 114; also Fig. 64) described and photographed this exposure as a good example of a "cliff of mylonite," one of the "excellent exposures of a fault breccia" on the campus of Rensselaer Polytechnic Institute. According to Ruedemann and reconfirmed by Elam (1960) a thrust fault follows part of this street (Sage Avenue) and Ruedemann mistook this conglomerate for a fault breccia. Perhaps the presence of criss-crossing veins in this exposure led to his interpretation of a "cliff of mylonite." Jack G. Elam (1960; unpublished Ph.D. thesis at Rensselaer Polytechnic Institute) assigned the rocks at this exposure to the Schodack lithofacies of Early Cambrian age. Cushing and Ruedemann (1914, p. 69) had introduced the "Schodack Formation" which according to Fisher (1961, p. D8) has now fallen victim to a nomenclatorial "snafu." Zen (1964) has renamed this formation the West Castleton Formation.

Lowman (1961) recognized that the boulders are a conglomerate and not a breccia, and following Kuenen and Migliorini (1950), he introduced the term <u>brecciolas</u>. The term <u>brecciolas</u> refers to graded limestone breccia beds that alternate with dark-colored shales (Lowman, 1961, p. B6; Sanders and Friedman, 1967, p. 242; Friedman, 1972, p. 25; Friedman and Sanders, 1978, p. 390,395).

The limestone- sandstone- and chert boulders which are embedded in shales at this exposure range from angular to rounded and show considerable variation in size (Fig. 6). Some boulders are coarse-grained fossiliferous limestone fragments with a micritic dolomite matrix. The rocks above the brecciolas are greenish-gray shales.

The boulders are those of rocks that formed under shallow shelf conditions. Their emplacement as boulders into shales, which are considered to be offshore deep-water sediment, indicates that the boulders moved downslope. The environment of deposition inferred for the brecciolas at this stop is that of the lower slope or base of slope (Fig. 1). Although the boulders came from the west down the slope, shelf carbonates (their source) extend many miles east of Troy (and presumably underlie Troy at depth).

Brecciolas which formed along the original east edge of the carbonate shelf parallel to the depositional strike for hundreds of miles define the site of the basin margin (rise) in Cambrian-Ordovician time. This Cambrian-Ordovician basin margin was located east of Troy near the present

Mileage Between points	Cumulative	

site of the Green Mountain axis. A relatively steep slope must have existed between the shelf edge and the basin margin with resultant instability that helped initiate slides, slumps, turbidity currents, mud flows, and sand falls.

		Continue uphill (east) on Sage Avenue, pass Student Union (on right) to Burdett Avenue, cross Burdett Avenue, and drive on to Parking Lot of Troy High School. Keep to extreme left (north end) of Parking Lot. Walk north to wall of old quarry.
0.4	87.5	STOP 6 TROY HIGH SCHOOL QUARRY

The spectacular brecciolas at this exposure consist of three members with eleven sub-members (Lowman, 1961). For details of the rocks, refer to Lowman's descriptions (1961, p. B11-B12). The brecciolas are Schodack lithofacies of the West Castleton Formation of Early Cambrian age, as at Stop 5.*

A thin-section study shows the limestones to consist of biomicrites, biointramicrites, and micrites with varying terrigenous quartz and clay minerals. The intraclasts are of pelmicrite. Shell fragments have been selectively dolomitized.

The observation that the limestone boulders are mostly micrites indicates that before removal downslope from their site of deposition the limestones were deposited under low-energy conditions on the shallow shelf to their original west, but at a place now still far to the east of Troy. The abundant fauna shows that the shallow waters were well aerated. The carbonate sediments must have lithified before their displacement downslope.

		Return to Burdett Avenue, turn right (north) on Burdett Avenue to Hoosick Street (NY7).
0.6	88.1	Turn right on Hoosick Street for one block and turn left on 21st Street to Troy Jewish Community Center.
0.3	88.4	STOP 7 TROY YMCA (in former Guidebooks [Friedman, 1972, 1979] listed as TROY JEWISH COMMUNITY CENTER)

*Note that the carbonate shelf west of Troy contains no rocks older than Late Cambrian; somewhere between west of Troy and the present Green Mountains the Middle Cambrian (?) and Lower Cambrian strata wedge in.

Mileage Between points	Cumulative

One large block of orthoquartzite, approximately 30 feet by 15 feet, probably settled in deep-water shale. Although limonitic, this orthoquartzite is devoid of rock fragments, hence is a second-cycle or multicycle rock. Note that the exposed shales surrounding this erratic block show that this block occurs singly. This exposure occurs along strike of the brecciolas and many more exposures of the brecciolas occur north of here in Frear Park. In a pit about 100 feet or so north of this block we exhumed from the shale a block of dark gray, fractured and veined micritic dolomitic limestone.

The size and shape of the block of orthoquartzite suggests more than a steep slope. To detach a block of this dimension required considerable instability near the shelf edge, such as severe shakes as occur during earthquakes. This block of rock differs in lithology from the brecciolas which we have seen at the previous two stops. In contrast to the flat limestone boulders of the previous stop this huge block with its irregular outline suggests that solid bedrock of sandstone was forcibly detached from the shelf edge or basin slope. By analogy with modern events, turbidity currents, slumps, mud flows, and slides are usually funneled through submarine canyons. Could it be that this block was part of the wall of a submarine canyon which became detached during one of the slides and was moved by gravity into the basin or basin margin?

The alternative interpretation would be to consider this block to have been caught up in fault movement. Indeed slicken-sides are present on this block. However, the lower exposed contact with the shale is depositional and not faulted. Because the orthoquartzite block occurs along strike with the other brecciolas, and a limestone block has been found about 100 feet away, the evidence suggests that emplacement was by gravity rather than by faulting.

Convince yourself that this block is not a glacial erratic.

The shales at this stop have been assigned to the Schodack lithofacies of the West Castleton Formation of Early Cambrian age; the displaced orthoquartzite block may be as old as Precambrian.

		Return on 21st Street to Hoosick Street (NY7).
0.2	88.6	Turn right on Hoosick Street (NY7) and immediately

Mileage Between points	Cumulative	
		left (one block) on Burdett Ave., continue on Burdett Ave.
0.8	89.4	Turn right on Tibbits Avenue to Brunswick Ave.
0.1	89.5	Turn left on Brunswick Avenue.
0.2	89.7	Turn left on Congress Street, bear right at fork, follow NY66 across Poestenkill Bridge to Linden Avenue (first street after bridge).
0.1	89.8	Turn right on Linden Avenue.
0.1	89.9	Walk down ravine to bottom of Poestenkill Falls and gorge.
		STOP 7A POESTENKILL FALLS AND GORGE (Not on Fig. 11)

STOP 7A. Deep-Water Facies: Basin and/or Basin Margin

The bottom of Poestenkill Falls exposes Cambrian-Ordovician shales as well as interbedded graywackes and shales. Although stratigraphy and structure in this exposure are complex, we take you here because of the beauty of the falls. Similar shales and graywackes are exposed in many parts of Troy and its environs.

The deep-water origin of the shales has been traditionally inferred from the abundant and diverse graptolite assemblages. The fine grain size that characterizes shales indicates a low-energy environment below wave base, but how far below wave base or whether wave base was 35 feet, 200 feet or in excess of 2,000 feet cannot be inferred on the basis of grain size alone.

In a climatic belt dominated on the shelf by calcium carbonate-secreting organisms, the deep-water sediments should likewise show a dominance of carbonates. Yet testing the shales and graywackes for calcium carbonate or searching for fragments of fossil shells consisting of calcium carbonate has been unsuccessful. The only possible inference is that the deep-water sediments originated below the compensation depths of calcium carbonate. Although this depth level may have varied during past geologic time, inferred depths of the basin or basin margin should be of the order between 7000 and 14,000 feet, a level at which rapid dissolution of calcium carbonate

takes place in the modern oceans. Large blocks of limestone, such as brecciolas, can survive at this depth as has been shown on the flanks of the Bermuda Apron where gravity displaced shallow-water shells occur at a depth in excess of 14,000 feet (Friedman, 1964).

Some of the beds contain gray to black shale streaks characteristic of turbulent mud flows.

The interfingering of shelf carbonates with deep water shales has been documented for many parts of the geologic section (Garrison and Fischer, 1969; Wilson, 1969; Friedman, Barzel, and Derin, 1971). As carbonate particles dissolve below the compensation level and a source for clastic particles is unavailable, the deep basin becomes "starved." The basin facies of the Taconic Sequence observed here is an example of a "starved" sequence.

Ruedemann (1930, p. 144-145) shows "Logan's line" (now known as Emmons' line; Rodgers, 1970), the classical thrust plane which places Cambrian over Ordovician rocks, to surface at this site. Elam (1960) concurs and places Lower Cambrian rocks (his Poesten lithofacies, now West Castleton Formation; Zen, 1964) in contact with Middle Ordovician rocks (the Austin Glen Member of the Normanskill Formation).

During the American Industrial Revolution numerous factories were clustered on the north slope of Poestenkill Gorge making cotton cloth and curry combs, barbed wire and buckwheat flour, machines, and much more. Today obscure moldering industrial ruins testify to this former busy activity. The last of the abandoned mill buildings tumbled into the stream in the fierce flood of 1938.

		Continue downhill on Linden Avenue to Spring Avenue.
0.3	90.2	Turn right on Spring Avenue to Canal Avenue.
0.1	90.3	Turn left on Canal Avenue.
0.3	90.6	Turn left on 5th Avenue to corner with Madison Street.
0.1	90.7	Walk old wagon road uphill to ruins of former buildings, site of former abandoned quarry.
		Look at exposure in old quarry (Rushor's Quarry).
		STOP 7B RUSHOR'S QUARRY, TROY (not on Fig. 11)

Stop 7B. Deep-Water Facies: Basin Margin

At this stop we see basin-margin sediments devoid of brecciolas. The

Mileage Between points	Cumulative

clue to the presence of a paleoslope are sole marks on the undersides of graywacke beds. These marks are infillings (molds) of depressions that formed in the soft bottom clays (now shales) as particles in turbidity current flows or sandfalls scoured or gouged the bottom. The following sole marks can be seen (definitions modified from Pettijohn and Potter, 1964):

flute molds - a raised sub-conical structure, the upcurrent end of which is rounded or bulbous, the other end flaring out and merging with the bedding plane;

groove molds - rounded or sharp-crested rectilinear ridges produced by filling of grooves;

brush marks - essentially a bounce cast with a crescentic depression on the down-current end;

prod molds - a short ridge, parallel to the current, which unlike flute molds, rises down-current, and ends abruptly;

frondescent marks - a type of load-flow structure that covers some soles with crowded lobate molds overlapping in the down-current direction.

The rocks in this quarry consist of interbedded graywacke and shale which show the characteristics of distal turbidites (Walker, 1967): cross laminae, convoluted laminae, sole marks, graded beds, parallel sides and regular beds, thin beds, fine grain size; individual sandstone beds rarely amalgamate.

The rocks at this exposure are part of the Austin Glen Member of the Normanskill Formation of Middle Ordovician age. In contrast to the Lower Cambrian (West Castleton) deep-water rocks which were tectonically emplaced in the Troy area, the Normanskill deep-water suite formed <u>in situ</u> after deep submergence of the Cambrian-Early Ordovician carbonate shelf.

		Return to Hoosick Street (NY7) and cross Hudson River over Collar City Bridge and proceed south on I787 to NY7 west; continue on NY7.
3.3	94.0	Latham Circle (continue on NY7)
0.5	94.5	Turn right (north); enter north entrance of Northway, Interstate 87.
1.9-2.0	96.4 to 96.5	Note two exposures of westernmost deep-water sedimentary facies, consisting of Middle Ordovician Normanskill graywacke and shale, in roadcut on right

Mileage Between points	Cumulative	
		(northbound lane). West of here and underneath the Normanskill at a depth of several thousand feet are Cambrian-Ordovician rocks of shelf facies.
0.4 to 0.5	96.9	Cross Mohawk River on Northway. This beautifully designed bridge won an award in 1958.
12.5	109.4	Note exposure of Middle Ordovician Canojoharie Shale, a dark gray silty shale of outer shelf to slope facies.
2.2	111.6	Take Exit 13N from Northway (sign: U.S.9 North Saratoga) and follow Route 9 north.
1.8	113.4	Note on left traffic light to Performing Arts Center (Main Gate to Saratoga Spa State Park), but continue straight for another 1.0 mile to traffic light for Performing Arts Center (sign: Saratoga Spa State Park, Summer Theater, Roosevelt Bath, Golf Course).
1.0	114.4	Turn left at traffic light and drive along pine- and spruce-lined lane through Saratoga Spa Golf Course.
1.0	115.4	Turn left (south) on NY50 and proceed 1,000 feet to East Parking Lot (necessitating another left turn). Park near box office of Arts Center.
		STOP SP SARATOGA SPA STATE PARK (not on Fig. 11).

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MOHAWK VALLEY EPISODIC DISCHARGES - THE GEOMORPHIC AND GLACIAL SEDIMENTARY RECORD

by

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INTRODUCTION

Our trip begins in Cohoes, at a place where one can be persuaded that the Mohawk is still a river. Hinckley dam, Erie Canal locks and control structures along the Mohawk, and hydroelectric power facilities on tributaries control to a large degree river discharge and sediment transport. The canal-river system is closely monitored and it floods locally, mainly in response to late winter ice jams. River level is dramatically lowered (about 12 feet at Schenectady) during winter canal closing. The Mohawk is spectacular only in April during spring thaw when the falls, by eastern standards, resemble Niagara. It is little wonder an Indian standing on the brink in pre-canal days was prompted to call the spot, woefully, Cohoes - "place of the broken canoe." Now, during the navigation season, most of the available discharge is used to operate the locks. In a dry summer, the dam above the Cohoes falls diverts the entire river into the Waterford locks and a neighboring hydrostation; the falls reduce to a trickle; and one may readily walk across the exposed bedrock gorge floor. If the Mohawk of today is a study in contrasts, its geological past is even more so. We will examine some of the products of varied depositional environments - subglacial, deep glacial lake, shoaling lake, nominal river discharge, and high-discharge, glacial lake outbursts (hlaups), between Cohoes and Little Falls (Figure 1).

PREGLACIAL MOHAWK VALLEY

Between Little Falls and Schenectady, the Mohawk River follows what appears to be its ancestral course. Through the eastern part of this reach, the rock gorge thalweg lies deep beneath the modern floodplain, e.g. 300' at Schenectady, 250' at Pattersonville, 130' at Tribes Hill, 120' at the "Noses." Southeast from Schenectady the preglacial gorge extends well below sea level and joins the Hudson gorge near Coeymans. East and north from Schenectady, the Mohawk's drainage routes are of late glacial age and Holocene (since ca. 14000 years before present. The lowland cut by the Mohawk is formed on block-faulted nearly flat-lying Cambrian and Ordovician carbonates, black shales and gray sandstones (Figure 2). Adirondack foothills and Appalachian Plateau scarps overlook the lowland from north and south.

Fig. 1. Physiography of east central New York. (From Landforms of New York, J.A. Bier)

GLACIAL HISTORY

The oldest exposed glacial deposits in the Mohawk Valley are tentatively assigned to the Mid-Wisconsinan (Hell Hollow and Ft. Johnson beds) (Figure 3). The Plum Point Interstade is represented in the valley by deep dissection of Hell Hollow tills and glacial lake clays that produced a landscape with minimum elevations along the valley axis close to what we see there now (Figure 4). During the Late Wisconsinan Nissouri Stade, ice again invaded and imponded the Mohawk Lowland from the northeast — emplaced the Mohawk till, locally over proglacial lake sediments ("black and tan") and drumlinized much of the landscape. During the ensuing Erie Interstade and Port Bruce Stade, ice cover remained in the eastern Mohawk (LaFleur, 1983) while farther west (Little Falls and beyond) the Mohawk lobe retreated and readvanced. Active ice from Oneida, Black

River, and Adirondack lobes combined to produce a varied stratigraphy of alternating tills, lake deposits and ice contact sediments (Muller, Franzi and Ridge, 1983).

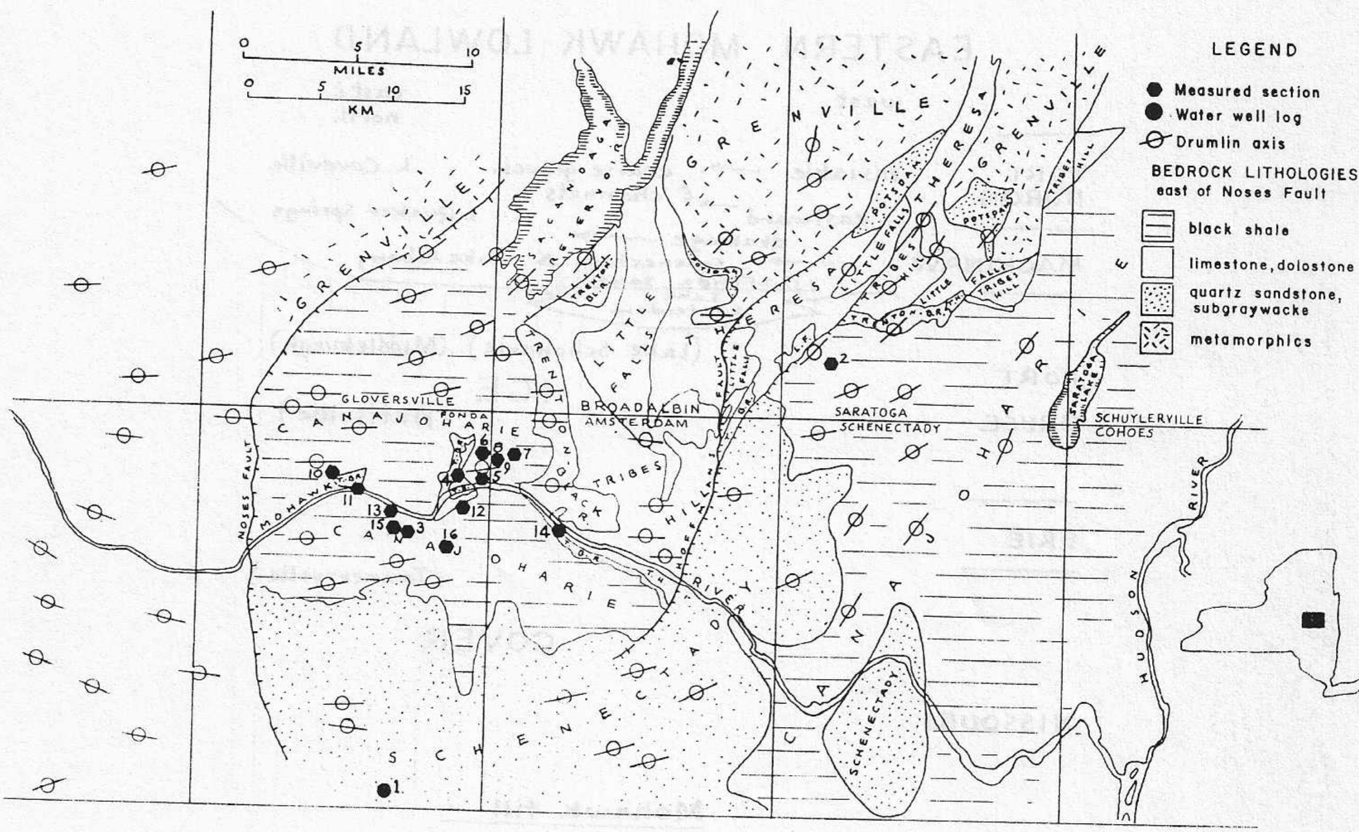


Fig. 2. Bedrock geologic map of eastern Mohawk Lowland showing location of measured glacial stratigraphic sections and drumlin axes. (Map after <u>Geologic Map of New York</u>, 1970)

Much of what we will see today took place in Late Port Bruce, Mackinaw, and Port Huron time, spanning perhaps 1500-2000 years, during which time the Mohawk valley east of Little Falls became deglaciated, was occupied by Lake Amsterdam and the Fonda wash plain, and then experienced a series of episodic high discharge events that may relate directly to glacial lake extinctions (hlaups) in the western Mohawk basin.

Meanwhile in the Hudson Lowland, Lakes Albany, Quaker Springs, Coveville, and Fort Ann formed, each at an elevation subordinate to its predecessor; each lake was impacted at a different place by the Mohawk River as it chose a variety of channels to reach the receding shores (LaFleur, 1975) (Hanson, 1977). Two questions arise:

1) to what extend did high discharge, catastrophic Mohawk hlaups influence the demise of each Hudson valley lake phase? 2) to what extent did channel routes of the Mohawk downstream from Schenectady change and relocate because of these same events?

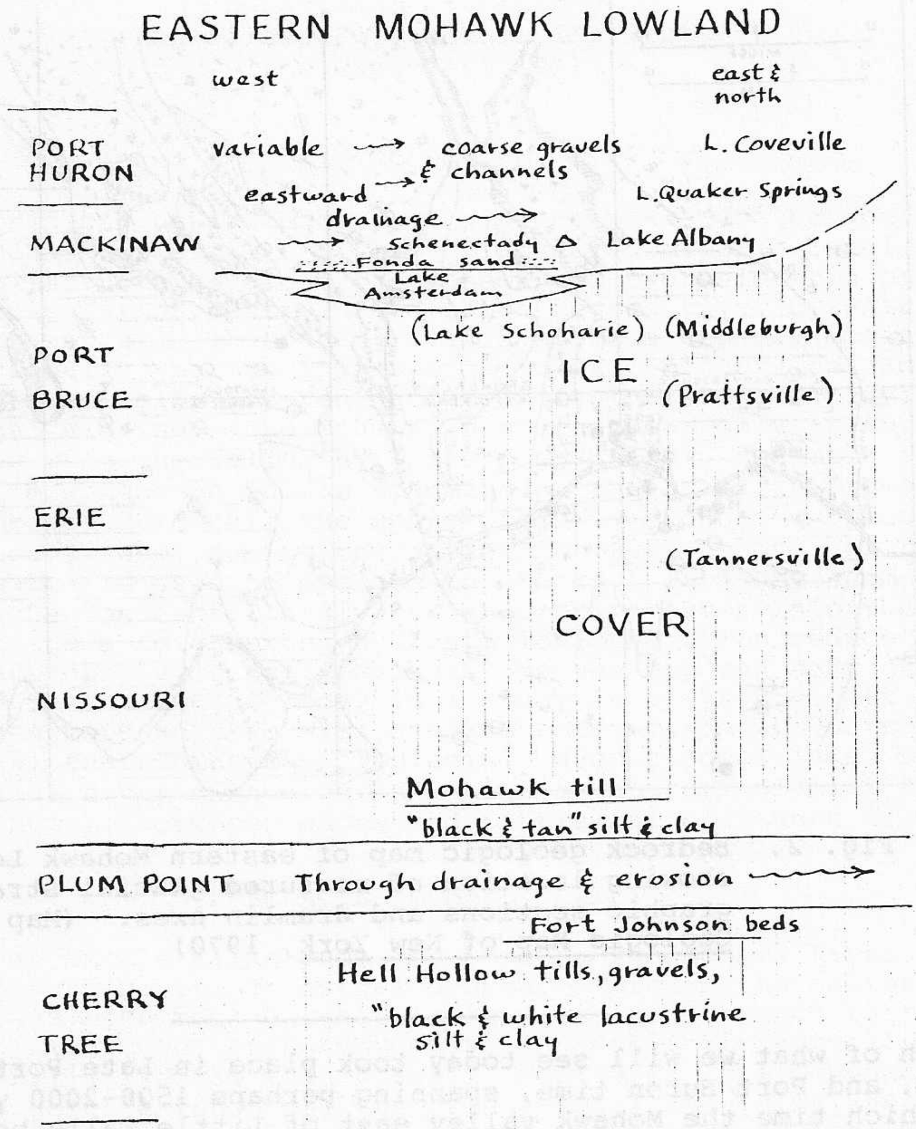


Fig. 3. Glacial stratigraphic synthesis of eastern Mohawk Lowland, on Middle and Late Wisconsinan time scale, spanning interval from about 35,000 to 12,500 years before present.

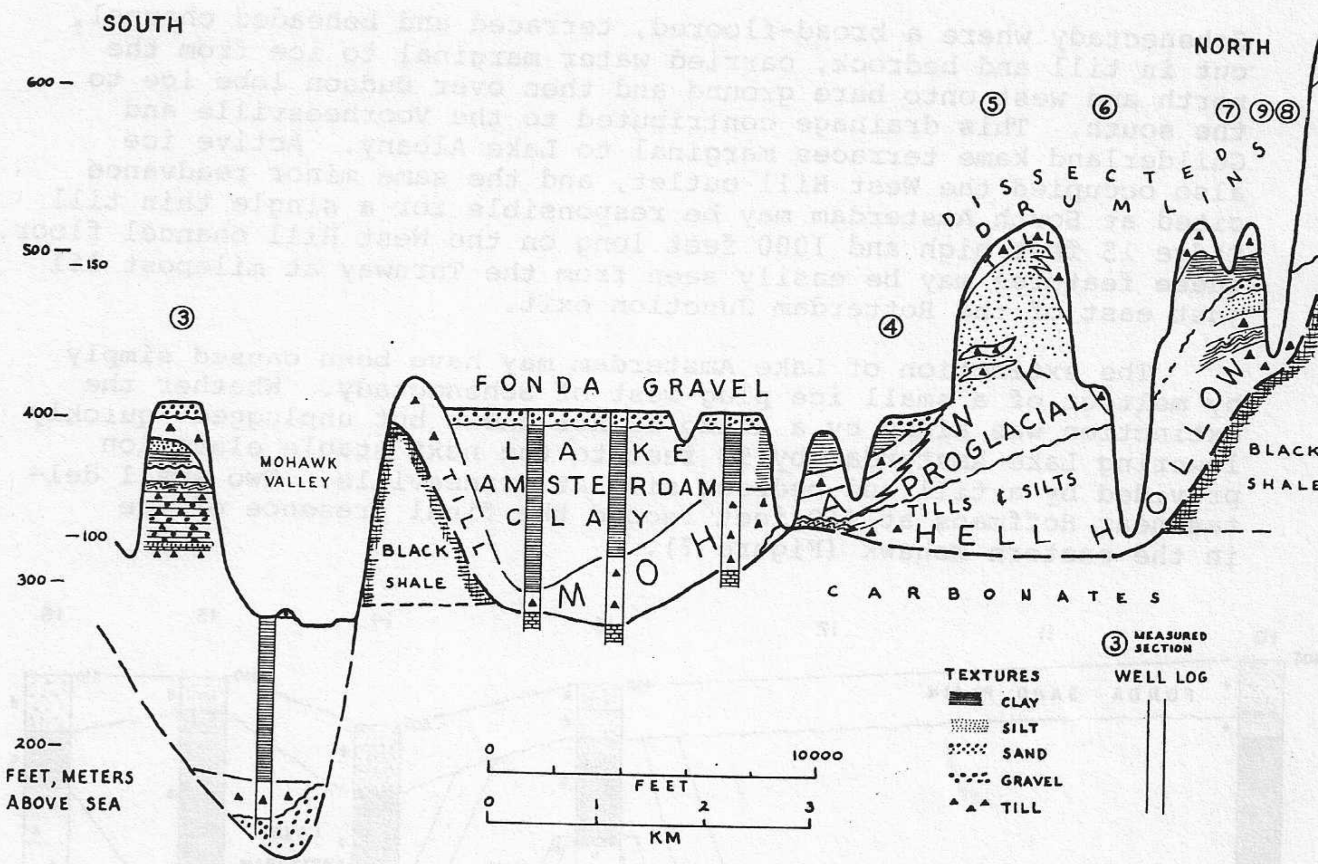


Fig. 4. Stratigraphic section across Mohawk Valley near Tribes Hill, facing west, based on measured sections, where numbered, and on water well records.

Lake Amsterdam

At many localities west of Amsterdam, Mohawk till is overlain by up to 50 feet of limy silt and clay rhythmites (Figure 5). There is no evidence that the Mohawk till was eroded by eastward-flowing water or was subaerially exposed prior to clay deposition. Lake Amsterdam accompanied deglaciation of the axial portion of the Mohawk Lowland while ice dammed its eastern end near Schenectady (Figure 6). Lake Amsterdam extended west at least to Little Falls and probably consisted of disjointed water bodies separated by isolated ice blocks and topographic highs. Although an ice surge of about one mile at South Amsterdam deformed and smoothed clay onto the stoss end of a Mohawk till drumlin, it does not appear that the lake was created by a readvance. The exit for overflowing Lake Amsterdam water is suggested at 500 feet at West Hill, west of

Schenectady where a broad-floored, terraced and beheaded channel, cut in till and bedrock, carried water marginal to ice from the north and west onto bare ground and then over Hudson lobe ice to the south. This drainage contributed to the Voorheesville and Guilderland kame terraces marginal to Lake Albany. Active ice also occupied the West Hill outlet, and the same minor readvance cited at South Amsterdam may be responsible for a single thin till ridge 15 feet high and 1000 feet long on the West Hill channel floor. These features may be easily seen from the Thruway at milepost 161 just east of the Rotterdam Junction exit.

The extinction of Lake Amsterdam may have been caused simply by meltout of a small ice plug west of Schenectady. Whether the extinction was aided by a hlaup is not known but unplugged quickly lowering Lake Amsterdam by 90 feet to the next stable elevation provided by a till and bedrock sill at Cranesville. Two small deltas near Hoffmans at 410 feet record the final presence of ice in the eastern Mohawk (Figure 7).

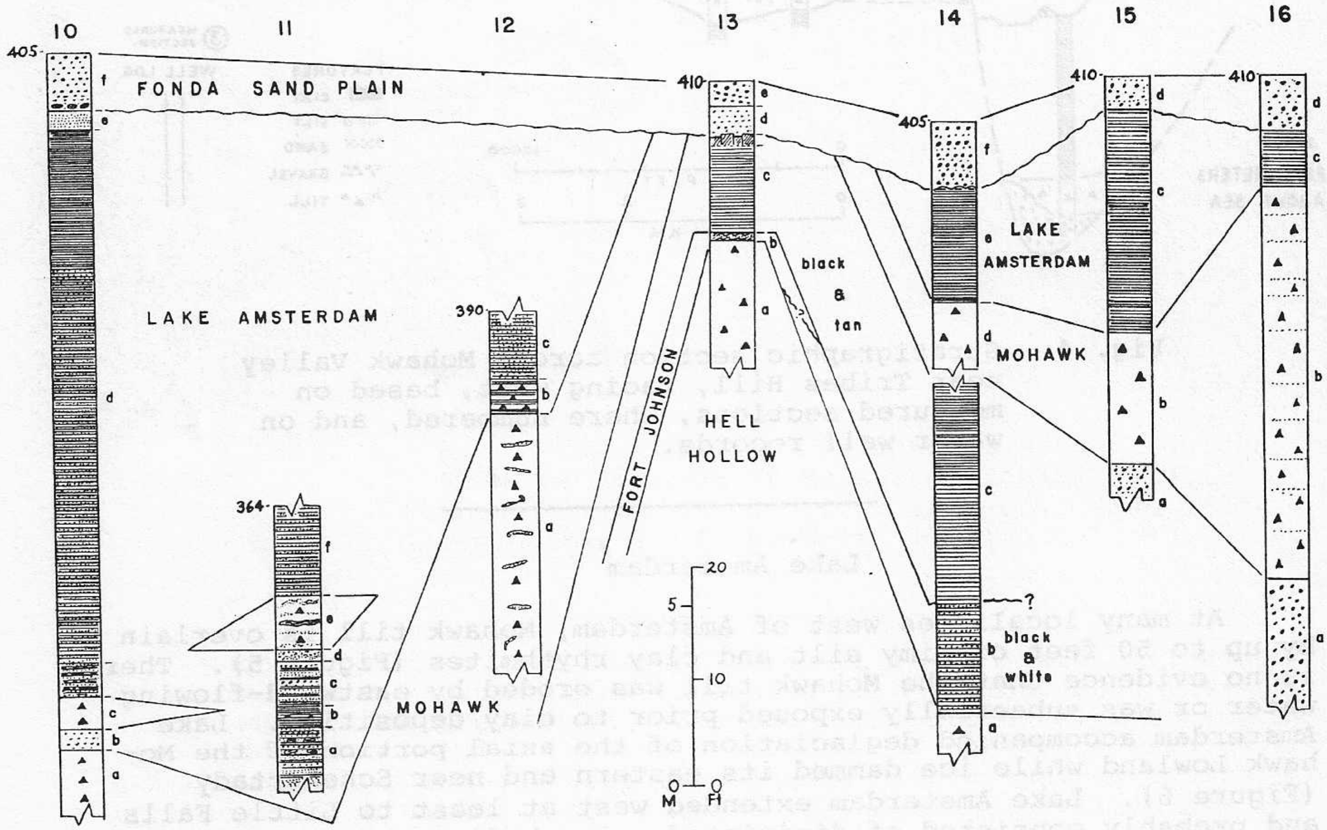


Fig. 5. Details of measured sections between Fonda and Amsterdam showing relationships between tills, glaciolacustrine, and wash plain sediments. Same legend as Figure 4 for textures. See Figure 2 for locations.

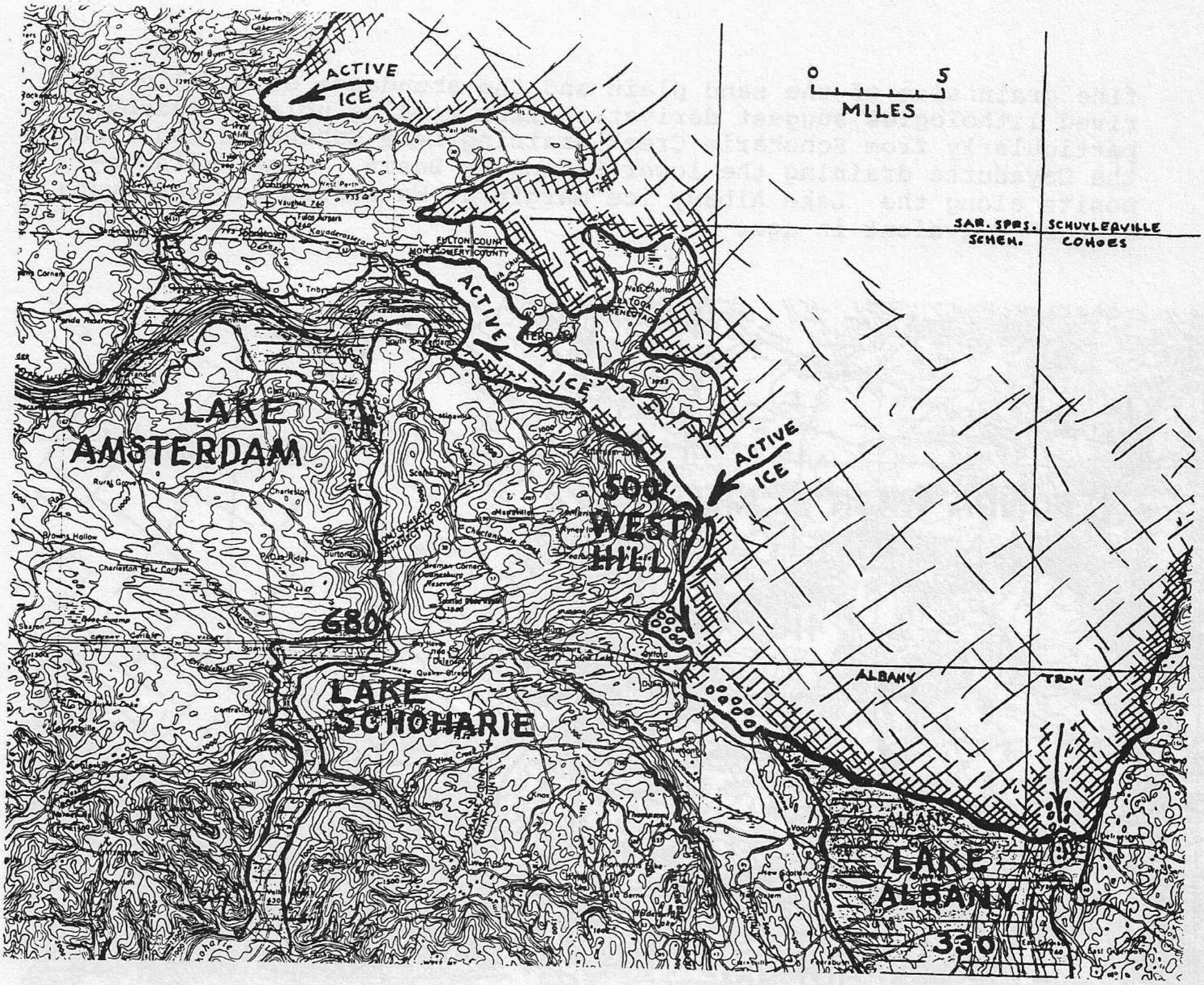


Fig. 6. Relationship of Hudson-Mohawk ice lobe border to glacial lakes Amsterdam and Albany. Only eastern half of Lake Amsterdam is shown. Numbers refer to elevation of outlets or lake levels in feet above sea.

Fonda Wash Plain

The Fonda wash plain at about 420-410 feet, recognized by Brigham (1929), consists of 5 to 10 feet of sand and pebble gravel overlying Lake Amsterdam clay (Figure 7). Sand terrace remnants can be traced westward from Cranesville to the mouth of East Canada Creek near St. Johnsville. The eastern Mohawk Valley was deglaciated by this time, but there was not accompanying high discharge from the west. The

fine grain size of the sand plain and the abundance of locally derived lithologies suggest derivation from eroding uplands nearby, particularly from Schoharie Creek draining Lake Schoharie and from the Cayadutta draining the lower Sacandaga Basin. Kame delta deposits along the Lake Albany ice margin at Waterford and Niskayuna appear equivalent in age.

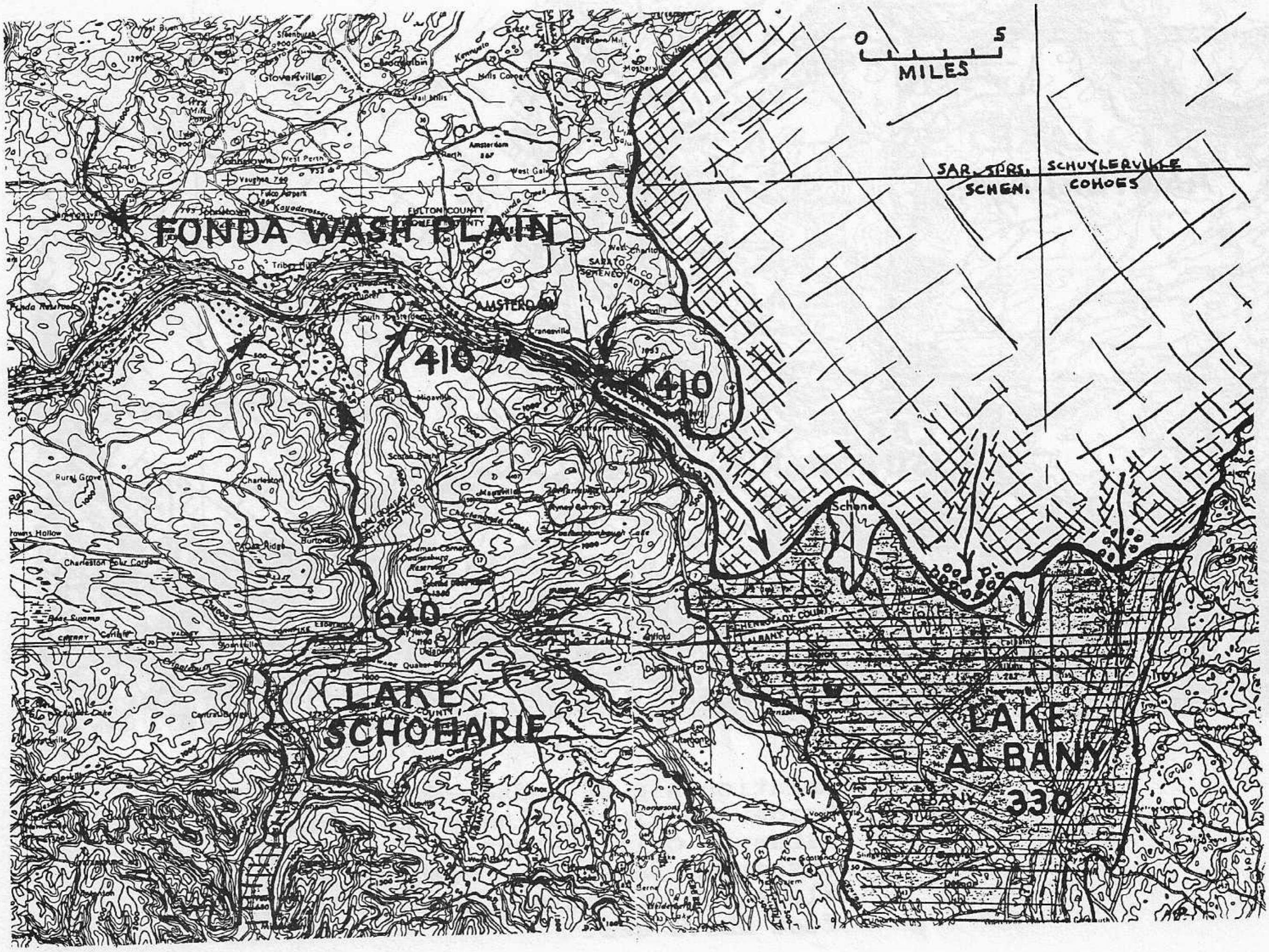


Fig. 7. Position of ice border just before unplugging of eastern end of Mohawk Valley. Fonda wash plain deposited by Mohawk tributaries over Lake Amsterdam clay.

Schenectady Delta

Northward recession of the Hudson lobe past Schenectady permitted the Mohawk to discharge directly into Lake Albany for the first

time (Figure 8). Dissection of the Fonda wash plain and Lake Amsterdam clay provided fine grained sediment for delta construction. Still this was not a time of abnormally high discharge through the Mohawk but rather a continuation of modest flow typical of the Fonda wash plain. North of Schenectady small deltas were built into Lake Albany at East Glenville by the Alplaus Kill, at that time an ice-margin stream. The Hoosick River constructed gravel terraces at 360 feet east of Schaghticoke, while at Halfmoon along the ice margin a kame delta complex formed in Lake Albany. The first of several detached ice blocks, all having importance in lake drainage history, was abandoned at Niskayuna.

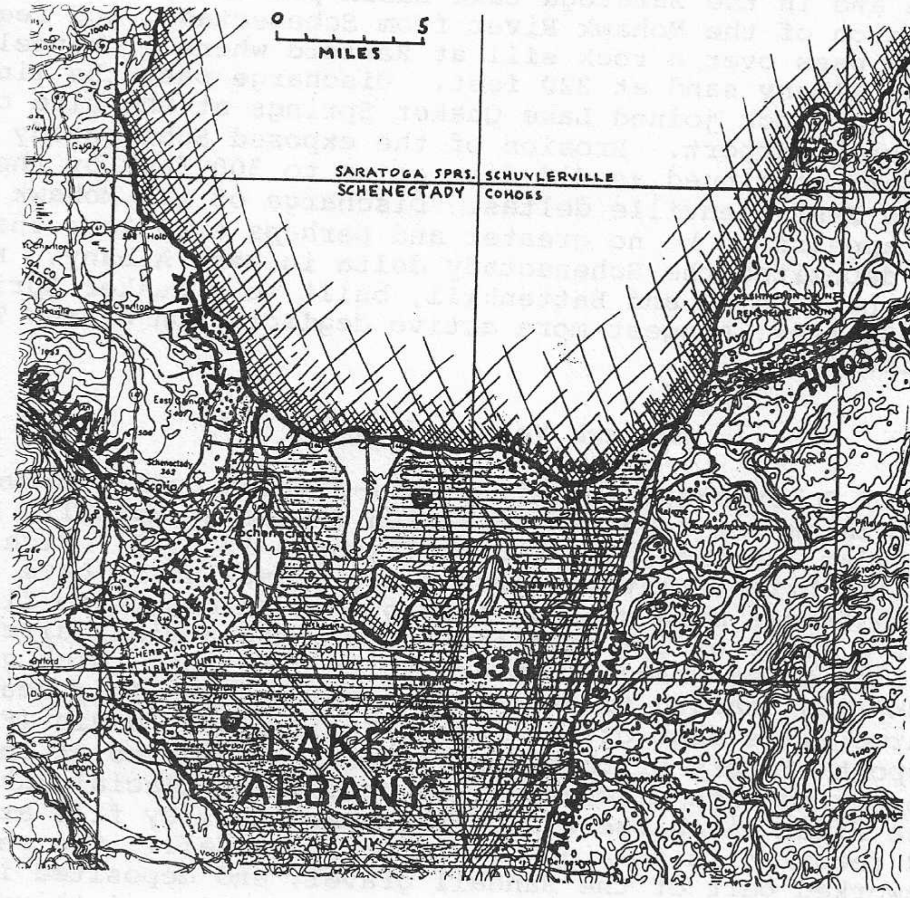


Fig. 8. Ice border position defending Lake Albany during early deposition of Schenectady delta.

Continued northward recession of the Hudson lobe margin toward Glens Falls was accompanied by lowering of Lake Albany (DeSimone, 1983). The reason for the drop in lake level is not apparent.

Discharge from the Mohawk continued to be modest through the duration of Lake Albany, and the Hudson lobe at this time shows little evidence of activity.

Lake Quaker Springs

Partial extent of the successor to Lake Albany, Lake Quaker Springs, some 50 feet lower, is indicated on Figure 9. This lake stage was defended on the north by ice receding through the upper Hudson and Champlain Lowlands. In addition to the ice block detached in Lake Albany near Niskayuna, others at Ballston Lake, Round Lake and in the Saratoga Lake Basin prevented immediate northward diversion of the Mohawk River from Schenectady, and required its flow to pass over a rock sill at Rexford where a channel was cut in Lake Albany sand at 320 feet. Discharge past the lingering Niskayuna ice block joined Lake Quaker Springs at the site of the present Albany airport. Erosion of the exposed Schenectady delta by the Mohawk continued as did dissection to 300 feet by the Alplaus Kill of the East Glenville deltas. Discharge of the Mohawk during this time appears to be no greater and perhaps even less than the flow that deposited the Schenectady delta in Lake Albany. Extensive deltas of the Hoosick and Battenkill, built to a Quaker Springs Lake level of 300 feet, suggest more active deglaciation of the Taconics than previously.

Mohawk Valley Gravels

West of Schenectady, massive, well-rounded cobble gravels up to 40 feet thick are found in separated masses at Randall, Fort Hunter, Rotterdam Junction, and Scotia (Figures 13, 17). Rich in western Mohawk basin "bright" lithologies, these gravel units have a grain size and thickness much greater than one might expect to find near the end of a river of low gradient (.0001) and nominal discharge. The source of these gravels appears to be a 50-foot-thick gravel unit which lies beneath tills, and is now exposed along Route NY5, one to three miles east of the Little Falls plunge basin (Figure 12). The transport history of these gravel masses requires at least three events involving high discharge from unconfined glacial lakes to the west. The first discharge eroded the gravel valley fill at Little Falls and reworked it to a new location at Randall. The second discharge reworked part of the Randall gravel, and deposited it as a valley fill between Rotterdam Junction and Scotia. A third discharge dissected the Scotia gravel and lowered the western half of the deposit to an elevation of 250 feet at Rotterdam Junction.

Summit elevation of the gravel mass at Randall lies at 350 feet, too low for correlation with Lake Albany. Lake Quaker Springs did not receive a Mohawk River delta of any consequence, so it would seem the first of these outbursts occurred late in Quaker Springs time, and may actually have caused the lake to lower.

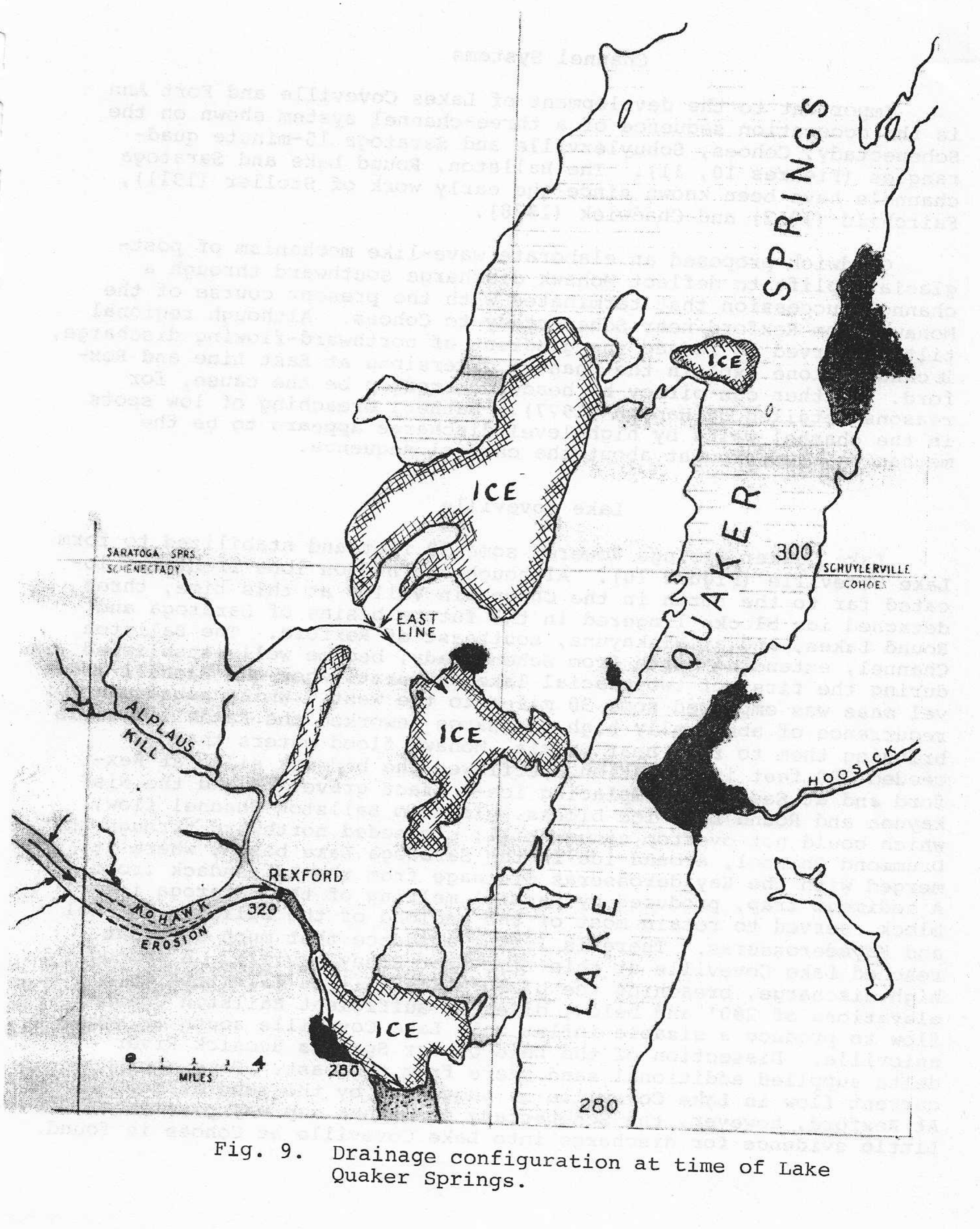


Fig. 9. Drainage configuration at time of Lake Quaker Springs.

Channel Systems

Important to the development of Lakes Coveville and Fort Ann is the occupation sequence of a three-channel system shown on the Schenectady, Cohoes, Schuylerville and Saratoga 15-minute quadrangles (Figures 10, 11). The Ballston, Round Lake and Saratoga channels have been known since the early work of Stoller (1911), Fairchild (1912) and Chadwick (1928).

Chadwick proposed an elaborate wave-like mechanism of post-glacial uplift to deflect Mohawk discharge southward through a channel succession that terminated with the present course of the Mohawk from Rexford near Schenectady to Cohoes. Although regional tilting served to reduce the gradient of northward-flowing discharge, it cannot alone explain the channel diversions at East Line and Rexford. Neither can piracy by headward erosion be the cause, for reasons detailed by Hanson (1977). Rather, breaching of low spots in the channel walls by high-level discharge appears to be the mechanism that brought about the channel sequence.

Lake Coveville

Lake Quaker Springs lowered some 50 feet and stabilized to form Lake Coveville (Figure 10). Although the Hudson lobe front was located far to the north in the Champlain Valley at this time, three detached ice blocks lingered in the future basins of Saratoga and Round Lakes, and at Niskayuna, southeast of Rexford. The Ballston Channel, extending north from Schenectady, became well established during the first of two glacial lake outbursts when the Randall gravel mass was emplaced some 30 miles to the west. While subsequent recurrence of abnormally high discharge reworked the Randall gravels bringing them to rest near Scotia, Mohawk flood waters that exceeded 300 feet in elevation overflowed the bedrock sills at Rexford and at East Line emplacing ice-contact gravel around the Niskayuna and Round Lake ice blocks. Interim Ballston Channel flow, which could not overtop either sill, proceeded northward through the Drummond Channel, around ice in the Saratoga Lake basin, where it merged with the Kayaderosseras drainage from the Adirondack front. A sediment trap, produced by partial melting of the Saratoga ice block, served to retain most of the bedload of the Ballston Channel and Kayaderosseras. There is little evidence that much sediment reached Lake Coveville at this time near Schuylerville. A second high discharge, breaching the black shale sill at East Line to elevations of 290' and below, diverted sufficient Ballston Channel flow to produce a sizable influx into Lake Coveville south of Mechanicville. Dissection of the Lake Quaker Springs Hoosick River delta supplied additional sand there from the east. Some southward current flow in Lake Coveville is suggested by the sand distribution. At Rexford, however, the Schenectady sandstone cap was persistent. Little evidence for discharge into Lake Coveville at Cohoes is found.

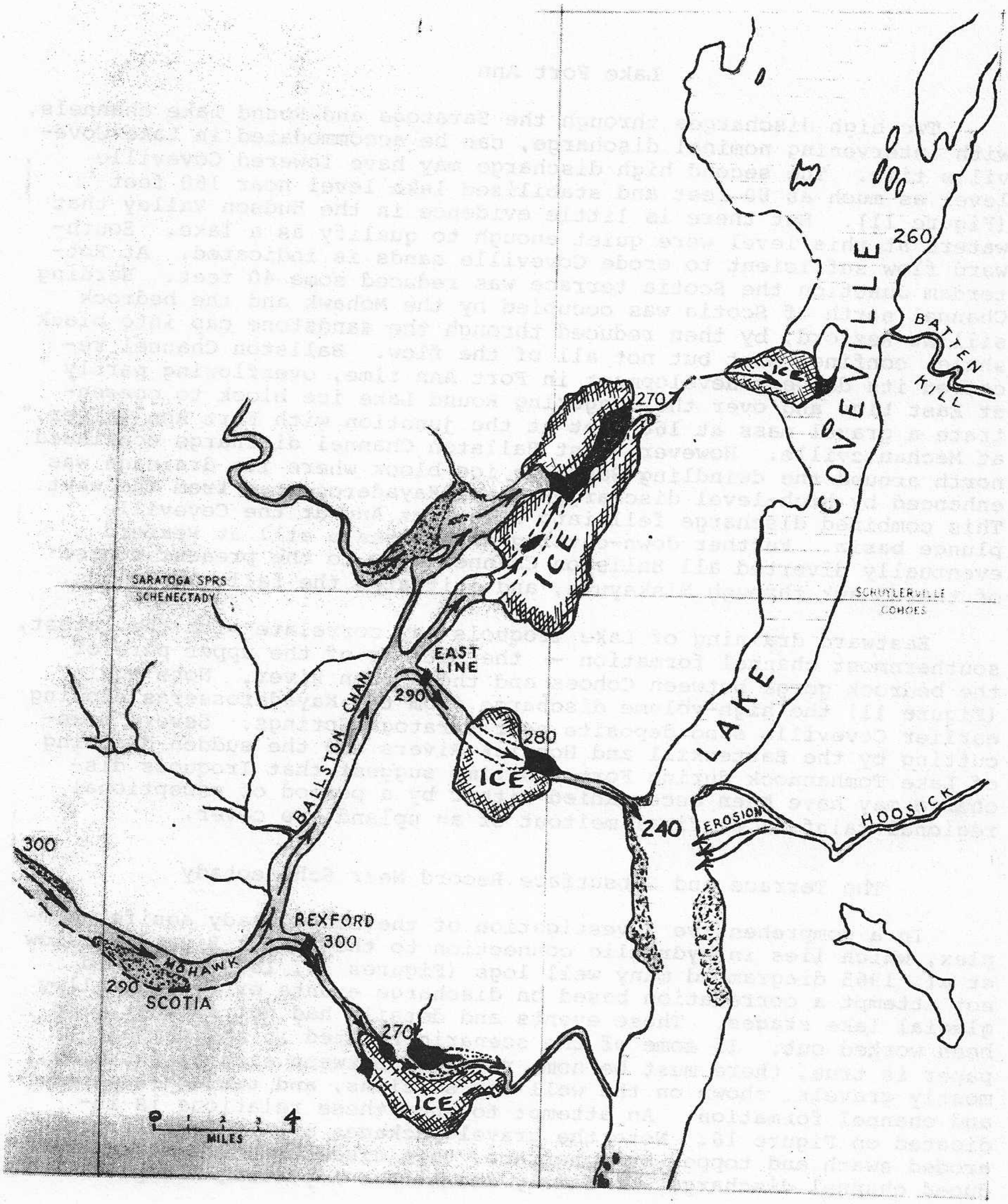


Fig. 10. Drainage configuration at time of Lake Coveville.

Lake Fort Ann

Two high discharges through the Saratoga and Round Lake channels, with intervening nominal discharge, can be accommodated in Lake Coveville time. The second high discharge may have lowered Coveville level as much as 80 feet and stabilized lake level near 160 feet (Figure 11). But there is little evidence in the Hudson Valley that waters at this level were quiet enough to qualify as a lake. Southward flow sufficient to erode Coveville sands is indicated. At Rotterdam Junction the Scotia terrace was reduced some 40 feet. Harding Channel north of Scotia was occupied by the Mohawk and the bedrock sill at Rexford, by then reduced through the sandstone cap into black shale, confined most but not all of the flow. Ballston Channel received its deepest development in Fort Ann time, overflowing partly at East Line and over the lingering Round Lake ice block to concentrate a gravel mass at 160 feet at the junction with Fort Ann "River," at Mechanicville. However, most Ballston Channel discharge continued north around the dwindling Saratoga ice block where its drainage was enhanced by high-level discharge of the Kayaderosseras from the west. This combined discharge fell into Lake Fort Ann at the Coveville plunge basin. Further down-cutting of the shale sill at Rexford eventually diverted all Ballston Channel flow to the present course of the Mohawk through Niskayuna, and initiated the falls at Cohoes.

Eastward draining of Lake Iroquois may correlate with the latest, southernmost channel formation — the cutting of the upper part of the bedrock gorge between Cohoes and the Hudson River. Note again (Figure 11) the high-volume discharge from the Kayaderosseras eroding earlier Coveville sand deposits near Saratoga Springs. Severe downcutting by the Battenkill and Hoosick Rivers and the sudden draining of Lake Tomhannock during Fort Ann time suggest that Iroquois discharge may have been accompanied either by a period of exceptional regional rainfall or final meltout of an upland ice cover.

The Terrace and Subsurface Record Near Schenectady

In a comprehensive investigation of the Schenectady aquifer complex, which lies in hydraulic connection to the Mohawk River, Winslow et al, 1965 diagrammed many well logs (Figures 14, 15), but could not attempt a correlation based on discharge events or Hudson valley glacial lake stages. These events and details had not at that time been worked out. If some of the scenario related in the present paper is true, there must be some relation between stratal packages, mostly gravels, shown on the well log sections, and terrace levels and channel formation. An attempt to show these relations is indicated on Figure 16. Note the gravel packages are emplaced in an eroded swath and topped by a terrace, then dissected by later, reduced channel discharge - a likely format for a passing hlaup.

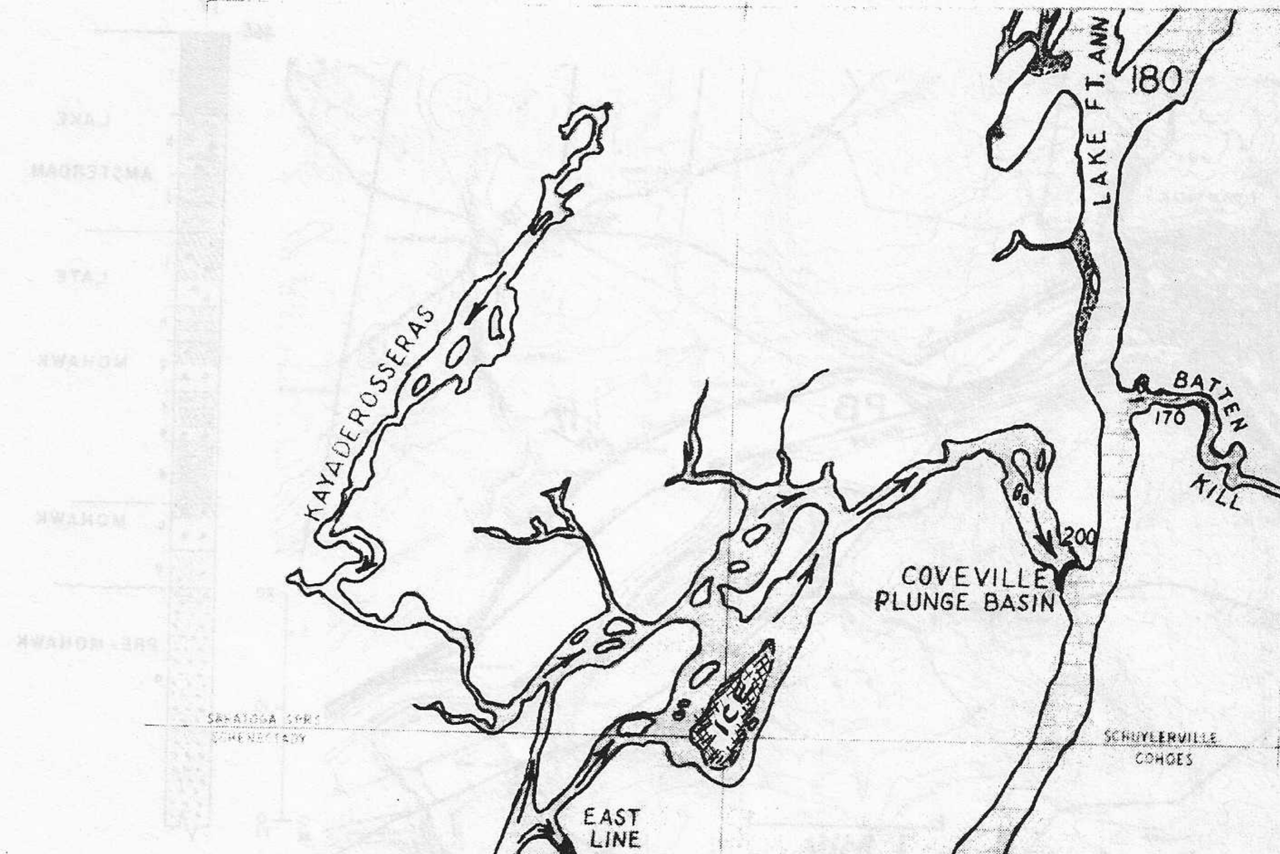


Fig. 11. Drainage configuration at time of Lake Ft. Ann.

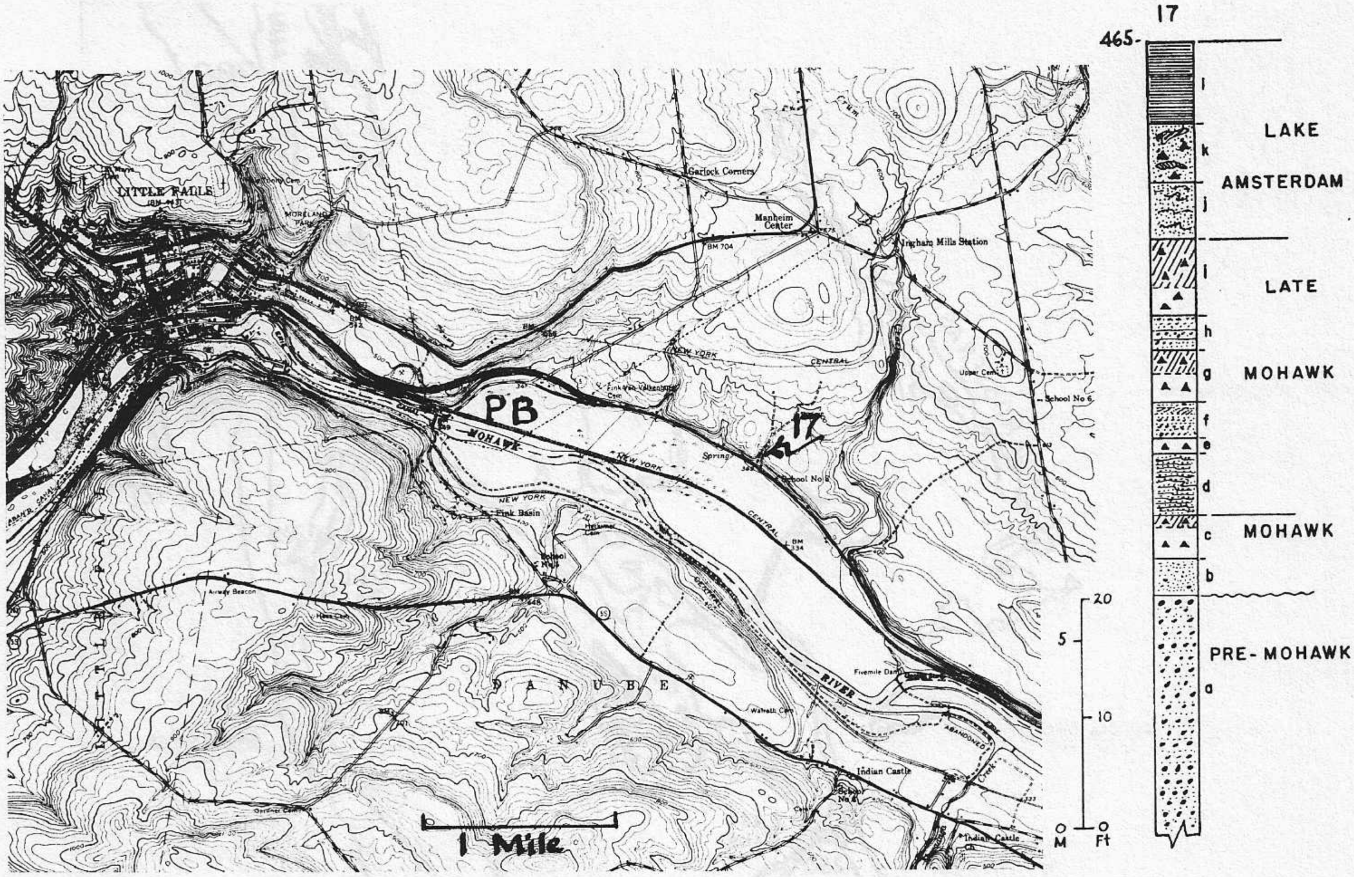


Fig. 12. Portion of the Little Falls 7.5' quadrangle, reduced in scale, showing Little Falls plunge basin, PB, and location of section 17.

Timing

We may now address the questions posed earlier. Coarse gravel redistribution in the eastern Mohawk Valley and the sequence of channels on the exposed Albany Lake plain can accommodate at least three episodes of late Woodfordian discharge through the Mohawk Valley including draining of Lake Iroquois.

Hlaups can be useful in lowering Hudson Valley lake stages because the sudden surcharge of floodwater can erode the ice or morainal dam that controls lake levels to the south. Three of these would suffice to cause transitions from Lake Quaker Springs to Coveville, and from Coveville through Lake Ft. Ann.

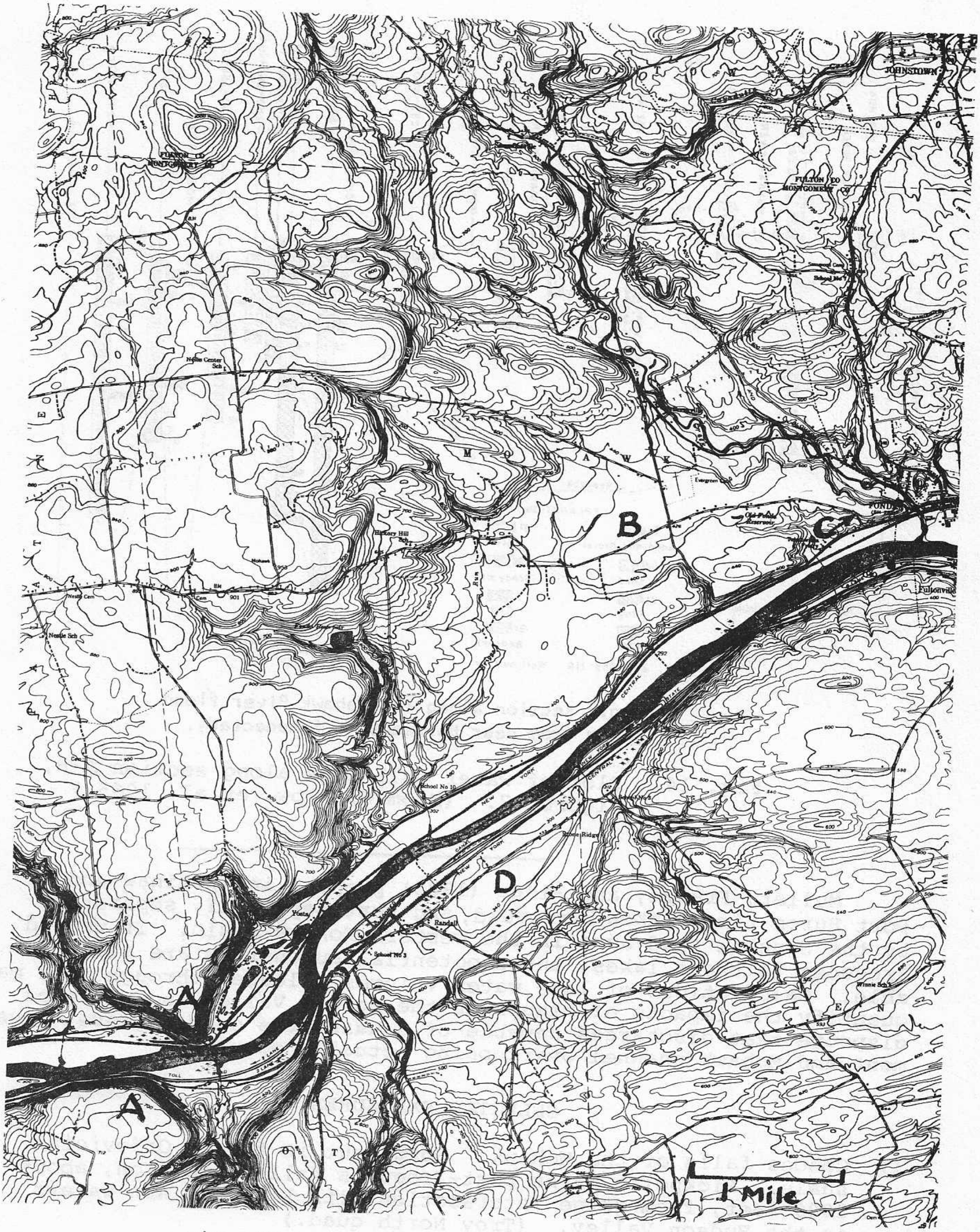
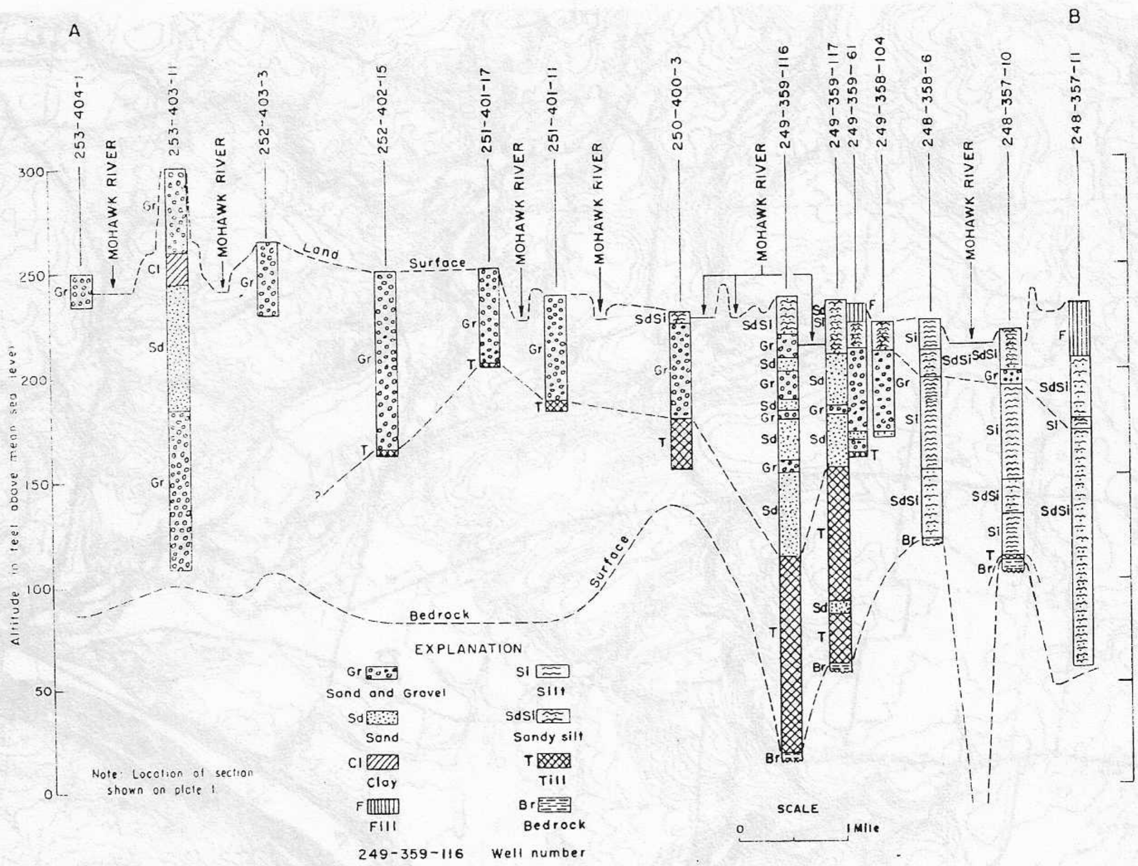


Fig. 13. Portion of Randall 7.5' quadrangle, reduced in scale, showing "The Noses," A; Fonda wash plain, B; Fonda/Amsterdam section 10, C; and Randall-Stone Ridge gravel mass, D.



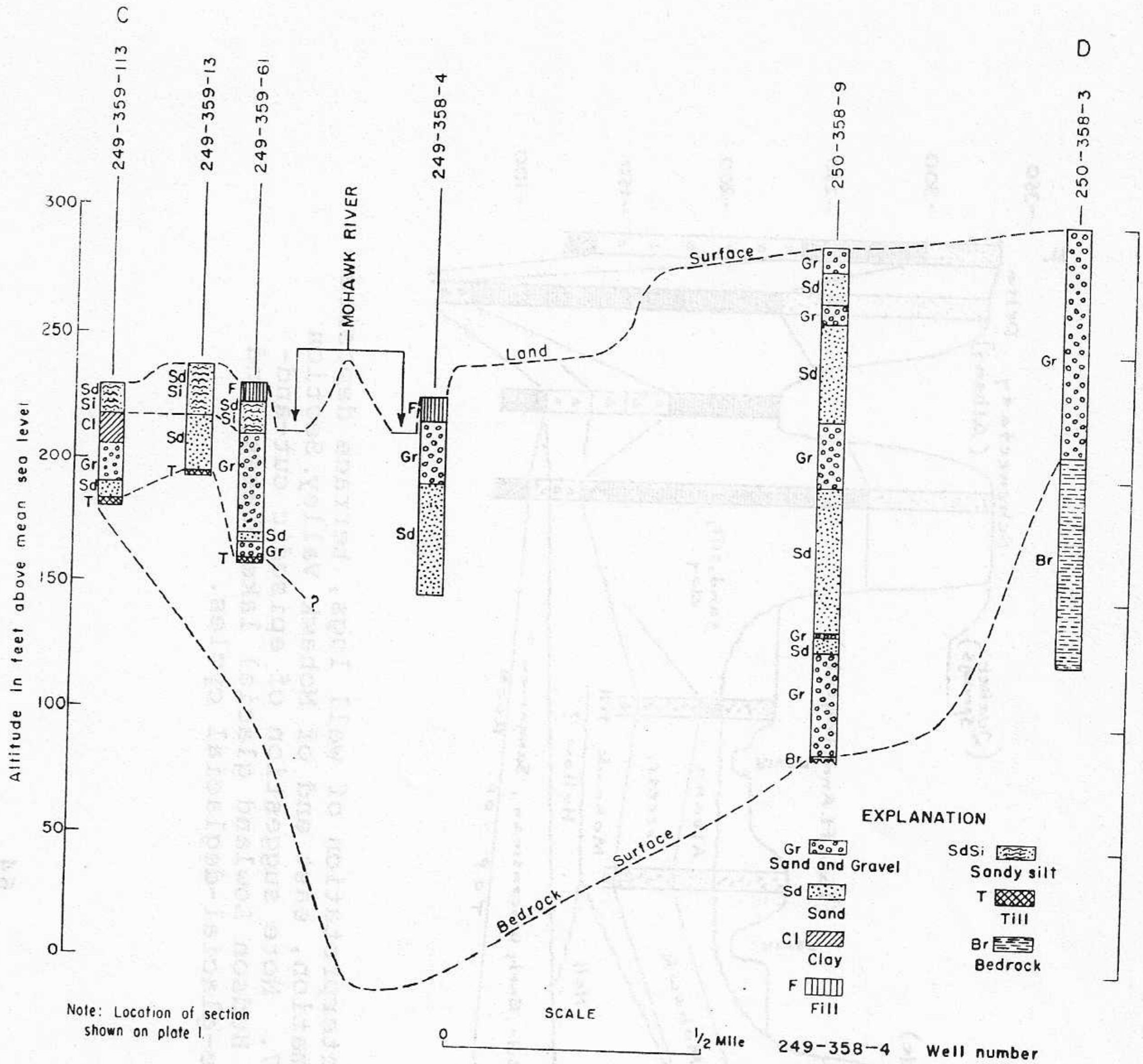
Geologic section along the Mohawk River flood plain from near Hoffmans to Schenectady.

Fig. 14. Plot of water well logs along section A-B, Figure 17. (From Winslow, et al, 1965.)

Muller, Franzi, and Ridge (1983) propose in Mackinaw and Port Huron time, Lakes Barneveld, and Port Leyden II, separated by the Stanwix readvance in the western Mohawk basin. If we add Iroquois to these lakes, three potential surcharges are available. Calculations of lake water volumes released through time may quantify the impacts of hlaups on the Mohawk Valley and prove or disprove that discharges capable of causing the coarse gravel movement and Mohawk channel overtopping actually occurred.

SELECTED EXPOSURES

1) Cohoes falls of the Mohawk. Gorge cut in folded Ordovician shale. Nickpoint marks post-Iroquois falls recession, about 2000 feet, from gorge intersection with Lake Ft. Ann waters in the Hudson Valley. (Troy North quad.)



Geologic section across the Mohawk valley from the Schenectady well field to Scotia.

Fig. 15. Plot of water well logs along section C-D, Figure 17. (From Winslow, et al, 1965.)

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2) Scotia terrace, 290' elev. Rt 147 (Sacandaga Rd.) Scotia. Gravel pit of Cushing Stone Co. exposes 30-35' of near-level-bedded cobble gravel of western Mohawk basin affinity; gravel mass is at least 2000 feet wide; terrace is one mile wide. (Lake Coveville equivalent) (Schenectady quad.)

3) Harding Crossing channel; at 250' parallels NYC tracks. Route of Mohawk prior to diversion to southwest side of Scotia terrace. (Lake Ft. Ann equiv.) (Schenectady quad.)

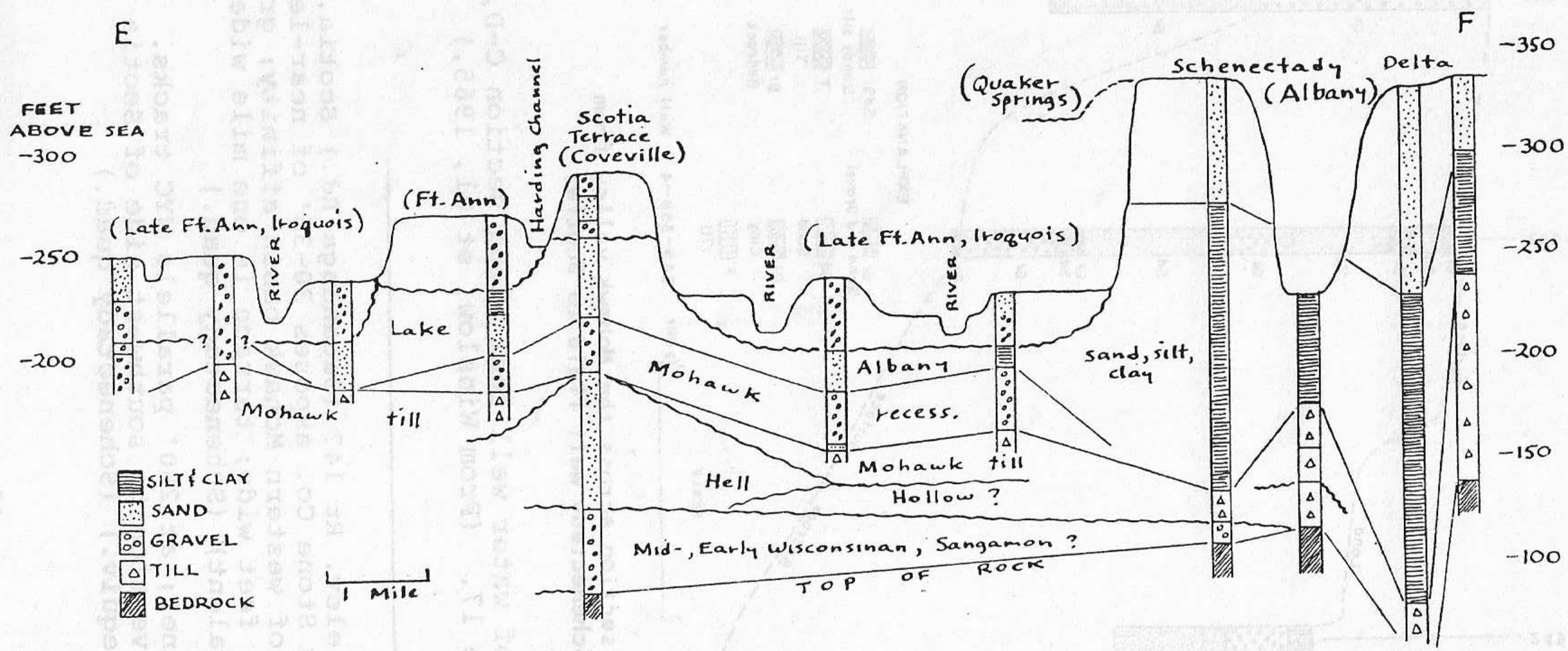

Fig. 16. Stratigraphic interpretation of well logs, terrace deposits and channel formation, east end of Mohawk Valley. Section E-F on Figure 17. Note suggestion of episodic cut-and-fill related to Hudson Lowland glacial lake phases, and deeper, multiple-glacial-deglacial cycles.

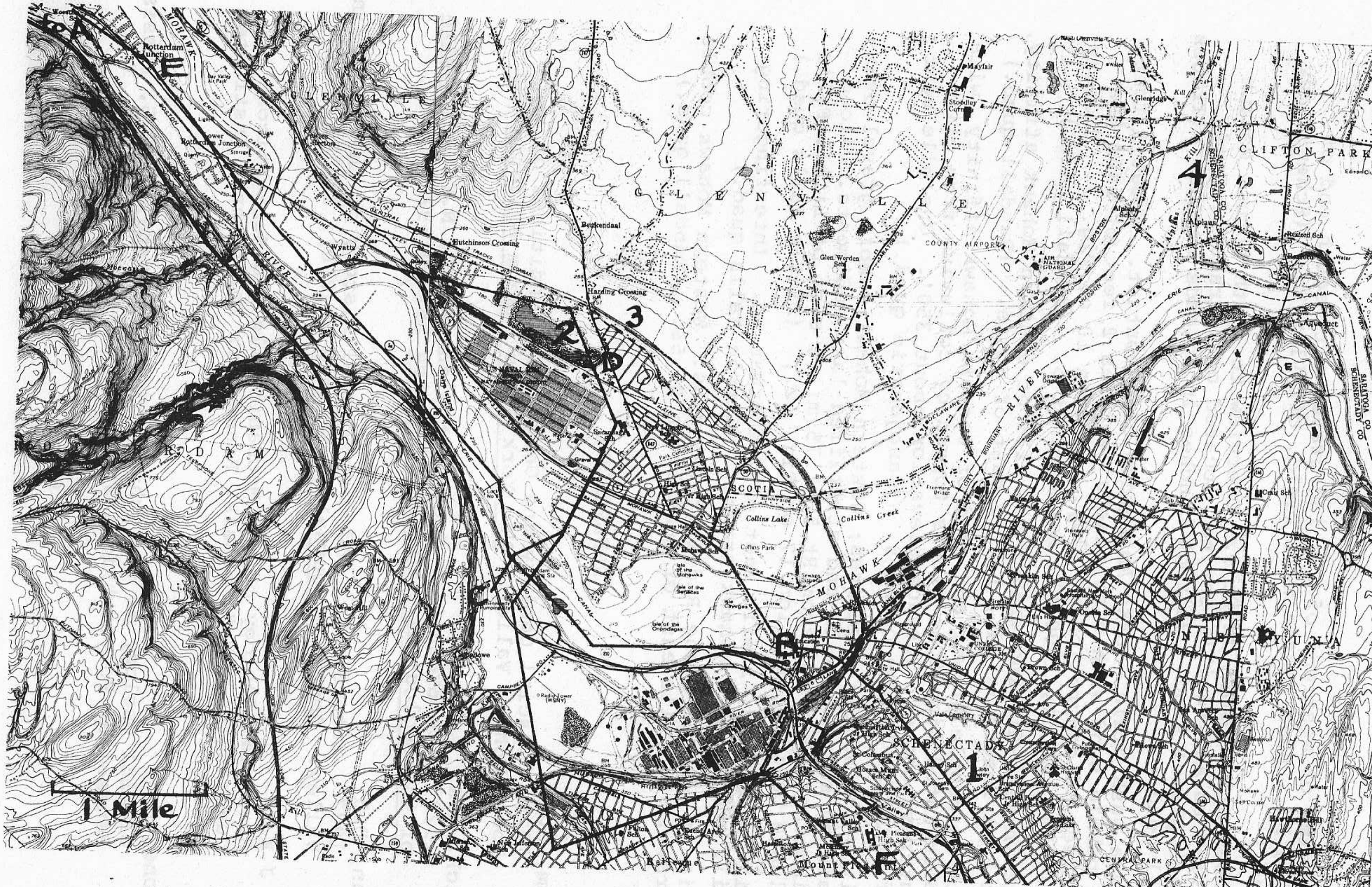

Fig. 17. Portions of the Rotterdam Junction and Schenectady 7.5' quadrangles, reduced in scale, showing locations of section lines A-B; C-D; E-F; Lake Albany Schenectady delta; Scotia terrace, 2; Harding channel 3; Alplaus channel, 4.

4) Overview of west end of Cushing gravel pit; south end of Ville Rd., near Scotia well field, off Vley Rd. (Schenectady quad.)

5) Tribes Hill summit on Rt 5. Overview to south across Mohawk Valley to Auriesville shrine, located on terrace of Fonda (wash plain) sand. Bedrock benches in uplands beyond. (Tribes Hill quad.)

6) Fonda wash plain - Lake Amsterdam clay contact. About 5 feet of pebble sand with clay clasts overlies 2 feet of sand and 50 feet of concretionary clay. Section 10 on Fig. 5. Hickory Hill Rd., Fonda, 1000 ft. north of Rt 5. (Randall quad.)

7) Pit in west-dipping gravel and till of eastern Mohawk affinity, Rt 5 at Reservoir Rd., 1 mile east of Yosts. (Randall quad.)

8) Pit in cobble gravel and sand of western Mohawk affinity, with colluviated limestone blocks topped by colluvium from cliff outcrops above. Terrace elevation about 350'. One mile east of Palatine Bridge on Rt 5. (Canajoharie quad.)

9) Pit and roadside exposures; western Mohawk cobble gravel overlain by several tills and stratified lake sediments. Along Timmerman Rd north from Rt 5, 1.75 miles east of Little Falls plunge basin. (Section 17 on Fig. 12.) (Little Falls quad.)

10) Pit on Ashe Rd, 1 mile east of Timmerman Rd. Interfingering till and western Mohawk gravel. (Little Falls quad.)

11) Pit in Randall-Stone Ridge gravel mass. Western Mohawk cobble gravel; 0.5 mile east of Randall on Rt 5S. (Randall quad.)

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